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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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PHOSPHORITE DEPOSITS ON THE SEA FLOOR OFF SOUTHERN CALIFORNIA

R. S. DIETZ, K. O. EMERY, AND F. P. SHEPARD

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ABSTRACT

Dredging operations on banks, on escarpments, and on walls of submarine canyons off southern California have shown that nodular phosphorite is the most abundant type of rock in these nondepositional environments. Approximately one-third of all the rock recovered is phosphorite. Petrographic and megascopic examination reveals that the nodules are largely formed by direct precipitation, but that they enclose some replaced material. Examination also shows that the phosphorite was probably deposited essentially *in situ*. Miocene Foraminifera have been identified in the nodules from many of the stations, but Recent or Pleistocene faunas have also been found in some of the phosphorite. Whereas the enclosed Miocene fauna suggests a Miocene age for most of the nodules, significant nonpaleontological data indicate that this fauna has probably been reworked into more recent deposits before enclosure in the nodules.

INTRODUCTION

During the series of geological cruises of the E. W. SCRIPPS in 1938, rock was dredged from a large number of stations off the southern California coast and from a few stations farther north. This rock will be described in a future paper by Emery and Shepard, but it was thought that special treatment should be given to the phosphorite because of its surprising abundance and because of the special problems which it raised. Phosphorite was recovered in greater abundance and was found in more localities than any other type of rock. Its abundance on the submarine surfaces in contrast to its scarcity in outcrops over any area of the same size on land, and the possibility that the present environment is

that in which the nodules were formed gave special interest to the investigation.

The large nodules were collected in bucket dredges (Shepard and Emery, 1940, Pl. 5, fig. 2). Many of the smaller ones were found in snapper samples and in cores of unconsolidated sediment.

Dietz made the principal laboratory study of the phosphorite samples but was assisted by Emery especially in the chemical investigation. Shepard conducted the marine investigation of the deposits. All of the writers have shared in the discussion of the origin of the material.

ACKNOWLEDGMENTS

The opportunity to collect samples from the sea floor was provided by a grant from the Penrose Bequest of the Geological Society of America. This grant made possible the use of the auxiliary schooner E. W. Scripps. The active co-operation of the Scripps Institution of Oceanography in field and laboratory has been important. Roger Revelle collected some of the samples and also assisted in a study of the solubility of calcium phosphate in sea water. The chemical data were checked by W. W. Rubey, W. L. Hill, K. D. Jacob, and W. F. Rodebush. For the age determination of Foraminifera of the phosphorite, the writers are greatly obligated to M. L. Natland of the Richfield Oil Company of California. Some Foraminifera were also identified by J. A. Cushman and S. G. Wissler. Critical readings of the manuscript were made by Eliot Blackwelder, M. I. Goldman, W. L. Hill, K. D. Jacob, G. R. Mansfield, W. W. Rubey, R. P. Sharp, H. R. Wanless, and R. C. Wells. Six chemical analyses of the phosphorite nodules were made by the U. S. Geological Survey. Laboratory aid was given by personnel of the Works Progress Administration, Official Project No. 665-07-3-141. The writers wish to express appreciation to these organizations and individuals.

DESCRIPTION OF THE PHOSPHORITE

GENERAL

About 600 kilograms of nodular phosphorite was dredged from 48 different stations, most of which are located off southern California (Pl. 3). The phosphorite ranges from oölites to large nodules (Pl. 1, fig. 4) measuring up to a maximum of 60 by 50 by 20 centimeters. Average diameter of the nodules is about 5 centimeters. Weights of the four largest nodules range from 35 to 73 kilograms.

EXTERNAL CHARACTER

The phosphorite varies in shape from thin flat slabs to nodular masses, but those from any particular dredge haul usually have a group resem-

blance. Smoothly rounded protuberances and cavities are present on some nodules (Pl. 1, fig. 1).

In general, the nodules are hard and dense and have a smooth glazed surface and a fresh appearance. Their surface consists of a thin discolored layer of phosphorite or of manganese oxide which obscures the internal character. Freshly broken surfaces are usually light to dark brown, but black surfaces are also found. At the four stations in the Santa Monica Bay area all of the phosphorite is black, both internally and externally.

In addition to occurring in nodular form, the phosphorite often coats hard compact rock and thoroughly permeates more porous varieties. Also, at several stations large pieces of conglomerate or breccia were dredged in which the phosphorite cements fragments of darker phosphorite and other types of rock (Pl. 1, fig. 2).

Many marine organisms including corals, bryozoans, sponges, serpulid worms, and brachiopods are found attached to the nodules. Some of the phosphorite is also deeply drilled by boring animals. In many specimens these borings and other cavities have become partially or completely filled with limy and phosphatic debris (Pl. 2, fig. 4). The presence of attached organisms growing on the top of a nodule indicates the position in which the nodule rested on the bottom. The upper surface is also sometimes identified by a thin black coating of manganese oxide which develops only on the exposed side, and indicates an interruption in phosphorite deposition. Such a coating is commonly present on other rocks dredged from the offshore steep slopes and banks.

INTERNAL CHARACTER

Phosphate minerals.—Petrographic examination of thin sections supplemented by chemical analyses show that collophane, composed of isotropic carbonate fluorapatite, is the principal mineral form in which the phosphate is present. Associated with the collophane is an anisotropic mineral which is probably francolite,[1] a carbonate fluorapatite. This mineral is present as a replacement of organic remains and other limy debris that was originally entrapped in the nodules. Francolite also exists as fibrous concentric rings in and around oölites, and as small grains disseminated throughout the collophane. Apparently it forms largely by the gradual crystallization of collophane and by the migration of some of the calcium phosphate to replace calcium carbonate.

[1] Although collophane and francolite are probably similar chemically, they differ optically, in type of crystallinity, and in other physical properties and, hence, were regarded as different minerals. However, the term collophane is a name for the microcrystalline equivalents of minerals of the apatite group including francolite. In most respects the difference between collophane and francolite is analogous to that between chalcedony and quartz.

FIGURE 1.

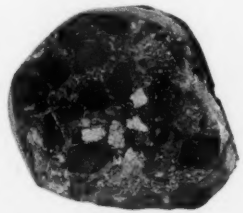


FIGURE 2.

FIGURE 3.

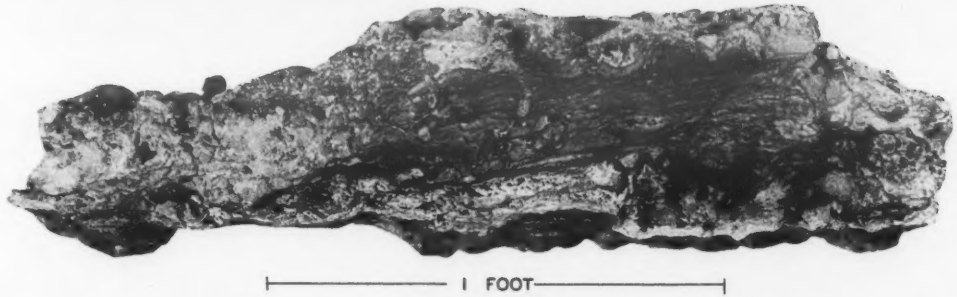


FIGURE 4.

FIGURE 1. NODULE FROM STATION 70
Showing the typical external appearance. Note the glazed surface and irregularly rounded shape. X ½

FIGURE 2. BRECCIA FROM STATION 63 CEMENTED WITH COLLOPHANE.
The fragments in the breccia are mainly phosphorite although some andesite and chlorite schist are also present. X ½

FIGURE 3. PHOSPHATIZED SEA LION BONE FROM STATION 48. X ½.

FIGURE 4. LAYERING ON FRESHLY BROKEN SURFACE OF NODULE FROM STATION 70. X ¼.

PHOSPHORITE NODULES AND PHOSPHATIZED BONE

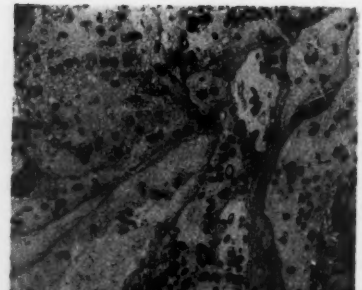
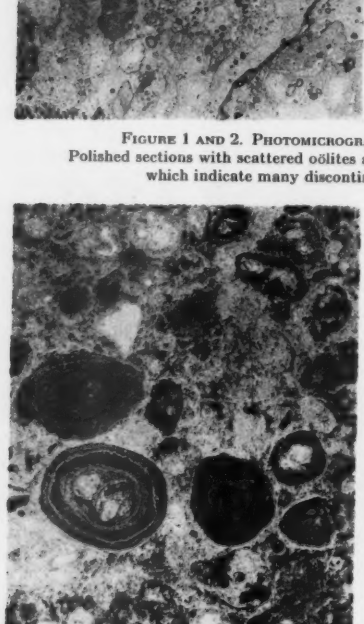


FIGURE 1 AND 2. PHOTOMICROGRAPHS OF NODULES FROM STATION 160
Polished sections with scattered oölites and mineral grains. Note the irregular layers which indicate many discontinuous periods of accretion. X 2½.



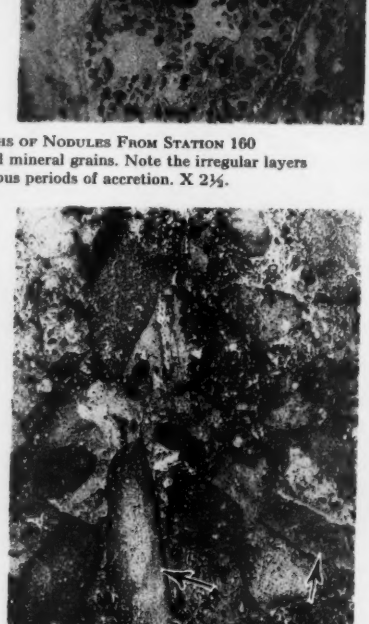


FIGURE 3. PHOTOMICROGRAPH OF A THIN SECTION OF NODULE FROM STATION 16
Under transmitted light shows typical texture. Note the concentric structure of the oölites. The nuclei consist of foraminiferal tests or mineral grains. X 28.

FIGURE 4. PHOTOMICROGRAPH OF A POLISHED SECTION OF A NODULE FROM STATION 162
Arrows point to animal borings which have been filled with sediment and subsequently phosphatized. X 3½.

Structure.—The structures of the nodules vary considerably in different samples and even within any particular haul. At a few stations there are pieces of homogeneous phosphorite composed entirely of massive collophane and francolite and having no enclosed material. However, the large majority of the nodules are layered and many are conglomeratic. The layers are irregular and nonconcentric (Pl. 2, figs. 1, 2), and vary in thickness from a few millimeters to a few centimeters. They are separated by a thin film of collophane of a darker shade of brown or less frequently by a thin coating of manganese oxide. Besides being layered, nearly all of the phosphorite is somewhat oölitic. However, oölites are much less abundant than in phosphorite samples from the Permian Phosphoria formation in which oölites form the mass of the rock.

Enclosed material.—A few phosphatic oölites are enclosed within the layers of phosphorite in most of the nodules (Pl. 2, figs. 1-3). The oölites are composed of concentric layers of collophane with an occasional layer of columnar francolite. Some of them have a definite nucleus (Pl. 2, fig. 3), which consists of a foraminiferal test, glauconite, or a clastic mineral grain. Usually, however, no nucleus can be discerned in thin section probably because the section failed to pass through the center of the oölite. In other instances, this absence is probably due to the complete absorption of a foraminiferal test nucleus. It is possible to find all stages in this absorption, from oölites having clearly defined tests of Foraminifera as nuclei to oölites in which only the faintest suggestion of the test remains.

Foraminiferal tests are also directly cemented into the nodules. Although a few tests are composed of unaltered calcium carbonate, most of them are more or less completely replaced by calcium phosphate usually in the form of francolite. In some nodules Foraminifera are extremely abundant and make up an appreciable part of the rock.

Although shells other than foraminiferal tests are not common within the phosphorite, fragments of corals, bryozoans, and other animal remains are found. At a number of stations, phosphatized bone fragments (Pl. 1, fig. 3) are present either separate or cemented into the nodules. These consist of bones of fish and mammals, the latter including a portion of a sea lion humerus and a whale's ear bone. At one station a phosphatized part of a turtle plastron was dredged. A number of other unidentified structures may be of organic origin.

Grains of clastic minerals are extremely abundant in some nodules but are absent in others. Some grains are rounded although more are angular. They consist mainly of feldspars, quartz, mica, and pyroxene.

Glauconite is abundant in nodules from most of the stations and occurs as rounded grains, as fillings of foraminiferal tests, as irregular masses, and as finely disseminated particles.

Insoluble residues.—To obtain insoluble residues, samples from many of the nodules were digested in hydrochloric or nitric acid. In typical samples this residue constitutes from 6 to 30 per cent, and averages 17 per cent.

The fine fraction of the residue consists dominantly of amorphous silica and carbonaceous material. Glauconite is extremely common in the coarse fraction of most samples, and in many cases predominates over all other types of material. As in the sediment associated with the nodules, much of the glauconite is coated with manganese oxide. In a few samples internal casts of foraminiferal tests are composed of glauconite or amorphous silica. Many of these casts preserve even the fine details of the tests and, consequently, were extremely helpful in studying the Foraminifera.

Also present in the coarse fraction of the insoluble residue are mineral grains, carbonaceous particles, and siliceous animal remains. The siliceous remains consist dominantly of sponge spicules, but include some diatoms and Radiolaria. A few samples of dark-colored phosphorite contain hollow oval carbonaceous casts and small (about 1 mm. in diameter) round oval carbonaceous pellets which may be coprolites.

ASSOCIATED SEDIMENTS

Examination of a large number of samples has shown that the sediments of the continental borderland [2] off southern California (Pl. 3) can be classified into four types: (1) shelf sediments consisting of sand and silt with abundant mica; (2) near-shore basin sediments consisting of micaceous silty clays and occasional sandy lenses; (3) offshore basin sediments consisting of fine clays containing abundant tests of pelagic Foraminifera; and (4) offshore bank sediments consisting of sand-size material largely nonclastic and composed of glauconite, foraminiferal tests, shells, and some detrital mineral grains, and rock fragments. This last type is found also on some steep slopes and in a few localities on the shelf and in submarine canyons.

The glauconitic sediment (type 4) distributed in patches over rocky bottom, indicates that these are essentially areas of nondeposition. Al-

[2] Continental borderland is a term applied to the rugged submarine topography of great relief lying between a coast and the continental slope which leads to the oceanic abyss. The term shelf is more properly restricted to the relatively flat zone of sea floor, mostly less than 400 feet deep, which rims the land.

though the sand is variable, a typical sample proved to be 60 per cent glauconite, 25 per cent foraminiferal tests and other animal remains, 10 per cent grains and oölites of phosphorite, and 5 per cent clastic minerals. Glauconite in the sands associated with the phosphorite has

TABLE 1.—*Chemical analyses*

	Station					
	69	106	127	158	162	183
CaO	47.35	45.43	45.52	46.58	37.19	47.41
R_2O_3	0.43	0.30	2.03	0.70	3.93	1.40
P_2O_5	29.56	29.19	28.96	29.09	22.43	29.66
CO_2	3.91	4.01	4.30	4.54	4.63	4.87
F	3.31	3.12	3.07	3.15	2.47	3.36
Org.	0.10	1.90	2.25	0.44	0.35	1.50
Insol in HCl	2.59	3.57	4.45	3.57	20.99	2.12
Total	87.25	87.52	90.58	88.07	91.99	90.32
F/P_2O_5	.1120	.1069	.1060	.1082	.1099	.1132
CO_2/P_2O_5	.1323	.1373	.1486	.1560	.2060	.1641

Analyses (excepting CO_2 and CaO) by U. S. Geological Survey.
CO_2 and CaO analyses by McVicker, Illinois State Geological Survey.

a dark green to black surface, the latter due to a coating of manganese oxide. Many of the grains are oval pellets or casts of foraminiferal tests. The composition of these sediments corresponds closely to material enclosed in the phosphorite nodules. Furthermore, phosphorite is almost invariably found associated with this type of sediment.

The association of phosphorite nodules with such deposits was also found by the Challenger expedition (Murray and Renard, 1891, p. 395-397). Furthermore, geologists have noted the common association of phosphorite nodules and glauconite in nondepositional horizons in sedimentary rock (Goldman, 1922).

CHEMICAL AND X-RAY INVESTIGATIONS OF THE PHOSPHORITE

CHEMICAL ANALYSES

Chemical analyses of nodules from six different stations were made by the U. S. Geological Survey. These analyses (Table 1) are surpris-

ingly uniform with the exception of sample 162 which differs mainly in the large amount of material insoluble in acid. The difference between the total percentage of the analyses and 100 per cent is probably largely MgO and soluble SiO_2 although small amounts of other substances must also be present.

The high fluorine content is of especial significance because it shows that the phosphate minerals are fluorapatites rather than hydroxyapatites. The amount of fluorine is slightly in excess of that required to completely fill with fluorine all the (OH, F) positions in the general apatite formula (Hendricks, *et al.*, 1931). This excess of fluorine is probably due to the replacement of some of the O positions by F rather than to the presence of a distinct fluorine mineral such as fluorite. Although fluorite has been detected in microscopic study of some phosphorites, for example the Phosphoria, fluorite was not found in the phosphorite from the sea floor off California.

In order to determine if the basin areas are the sites of unusually rapid deposition of phosphate, a series of phosphate analyses of muds from these environments was made by J. P. Cunningham. Samples were selected from the top and from the bottom of cores taken in basins at various depths and at different distances from shore. Phosphate varies from 0.3 to 0.5 per cent and shows no correlation with depth or distance from shore. The phosphate content is of the same order of magnitude as in the Carnegie samples which were dredged from the North and South Pacific (R. Revelle, personal communication), suggesting that only normal amounts of phosphate are found in the basin sediments.

X-RAY DATA

Although collophane is isotropic and, therefore has been considered amorphous, x-ray diffraction patterns show that the mineral is submicrocrystalline (Hendricks, *et al.*, 1931). X-ray studies by Dietz show that the phosphorite from off California, like other phosphorites, gives the typical apatite powder pattern. This apatitelike pattern is common to francolite and collophane.

In spite of the fact that calcium oxide (Table 1) is present in the phosphorite from the California sea floor in fairly large quantities, calcite has not been detected by x-ray technique. Consequently, the calcium must be a component of the phosphate minerals.

Phosphorite from off California gives a pattern identical with that of a sample from the Phosphoria formation (Permian) at Garrison, Montana, except that the lines of the latter are not as broad. This indicates that the material from the Phosphoria is composed of crystals having

a larger particle size. If the samples are representative of the deposits from which they were obtained, this comparison may illustrate the well-known tendency for amorphous and submicrocrystalline substances to become more coarsely crystalline after a lapse of time.

PALEONTOLOGICAL DATA

VERTEBRATE REMAINS

A number of phosphatized vertebrate remains such as teeth, vertebrae, the ear bone of a whale, and other bones, were enclosed or associated with phosphorite. Bones of mammals, probably marine, and of fish were well represented. A phosphatized sea lion bone (Pl. 1, fig. 3) was dredged along with about 200 phosphorite nodules at Station 48. This bone was identified by Dr. Remington Kellog as the lower end of the shaft of the right humerus of a sea lion which closely resembles though differs from the Recent *Eumetopias stelleri*. The geologic range of this species extends into part of the Pleistocene. According to Kellog the specimen differs from the bones of a related Miocene (Temblor) sea lion although it may possibly fall within the range of variation of this species. Other bones were too fragmentary for age determination.

FORAMINIFERA

General.—All significant paleontological data regarding age were obtained from a study of Foraminifera. The identification of these Foraminifera has been almost entirely the work of M. L. Natland, although S. G. Wissler and J. A. Cushman have also made certain identifications.

Fauna of the associated sediment.—The sediment associated with the nodules contains abundant Foraminifera and in many cases it is predominantly composed of these organisms. Not only Recent Foraminifera are present but also many reworked fossil types, predominantly Miocene, which are often worn, fragmentary, and partially dissolved. At Station 70, almost half of the Foraminifera brushed off the surface of the nodules were Miocene forms. In contrast, mud in cores from the basins in this area rarely contain reworked fossils.

According to Natland, the Recent faunas present in the associated sediments and the Recent (or Pleistocene) faunas detected in a few nodules are, in a general way, indicative of the depth at which they are now found.

Fauna of the associated nonphosphatic rocks.—A study of the rocks dredged from the ocean floor off southern California has disclosed the presence of abundant sedimentary rocks containing Foraminifera. These

organic remains show that the rocks are dominantly Miocene. A specimen of very slightly consolidated mudstone from Station 273 contained a mixed fauna with forms diagnostic of Saucesian, Luisian, Mohnian, Delmontian, and Repetto age (lower Miocene to lower Pliocene, Table 2).

TABLE 2.—*Stratigraphic sequence of faunal units*
(Adapted from Kleinpell, 1938, Fig. 14)

Periods		So. Calif. Faunal Units
Recent and Pleistocene		Recent, Pleistocene, and uppermost Pliocene
Pliocene		Pico
		Repetto
Miocene	Upper	Delmontian
		Mohnian
	Middle	Luisian
		Relizian
	Lower	Saucesian
		Zemorrian

Fauna of the phosphorite.—Many of the phosphorite nodules contain poorly preserved foraminiferal tests most of which are completely replaced by francolite. The rock usually splits through rather than around these tests, making it necessary to identify the Foraminifera from cross sections. However, insoluble residues of some of the nodules yielded well-preserved internal casts composed of glauconite and silica. Quaternary Foraminifera which are unaltered and in an excellent state of preservation were found in some of the nodules.

Diagnostic Foraminifera were found in the nodules from 15 stations. These stations include nearly all those at which a large number of nodules were dredged. The foraminiferal data are briefly summarized as follows:

STATION 16

Foraminifera include *Anomalina salinasensis* Kleinpell, and *Siphogenerina branneri* Bagg. Relizian or L. Luisian (m. Miocene).

STATION 45

Among the Foraminifera present are: *Baggina robusta* Kleinpell, and *Valvulineria californica* Cushman. Luisian (u. m. Miocene).

STATION 48

Foraminifera include *Bolivina californica* Cushman, *Bulimina uvigerinaformis* Cushman and Kleinpell, and *Pulvinulinella gyroidinaformis* Cushman and Goudkoff. Lower Mohnian (l. u. Miocene).

STATION 63

Cassidulina fauna of the Lomita Quarry type (Galloway and Wissler, 1927) is present in the matrix which also encloses numerous nodules which on the basis of lithology are probably Miocene. A characteristic form is *Cassidulina subglobosa* var. *quadrata* Cushman and Hughes. (u. Pliocene to Recent)

STATION 64

The phosphorite in this sample fills pholad borings in an andesite cobble. A *Cassidulina* fauna of the Lomita Quarry type includes *Cassidulina subglobosa* var. *quadrata* Cushman and Hughes, and *Cassidulina limbata* Cushman and Hughes. (u. Pliocene to Recent)

STATION 69

A boulder-size nodule from this station consists of five layers. The bottom layer is conglomeratic and contains in the matrix a rich and well-preserved *Cassidulina* fauna of the Lomita Quarry type. Foraminifera present include *Cassidulina californica* Cushman and Hughes, *C. reflexa* Galloway and Wissler, *C. tortuosa* Cushman and Hughes, *C. limbata* Cushman and Hughes, and *C. subglobosa* var. *quadrata* Cushman and Hughes. This fauna is uppermost Pliocene to Recent. Small nodules enclosed in the matrix contain Luisian or Relizian (m. Miocene) Foraminifera: *Baggina* cf. *robusta* Kleinpell and a Miocene type of *Bolivina*. The upper four layers of this nodule are dark brown to black and contain a few poorly preserved Foraminifera of upper or middle Miocene (*Baggina robusta* Kleinpell and a Miocene *Valvulineria*). Inside this nodule are a number of pockets which have no exterior connection. Foraminifera in unconsolidated sediment in these pockets are Pleistocene or Recent. This fauna, however, differs from the Recent fauna scraped from the surface of the nodule.

STATION 70

Nodules from this station contain a middle and upper Miocene fauna including *Baggina californica* Cushman, *Bolivina imbricata* Cushman, *Bolivina sinuata* Galloway and Wissler, *Cassidulina crassa* d'Orbigny, *Planulina ornata* d'Orbigny, *Pullenia miocenica* Kleinpell, and *Uvigerinella* cf. *obesa* Cushman. Although the fauna in the sediment clinging to the surface of the nodules is predominantly Recent, an abundance

of reworked Miocene Foraminifera were found including *Siphogenerina collomi* Cushman, *Cassidulina monicana* Cushman and Kleinpell, and *Uvigerina hootsi* Rankin.

STATION 106

The largest nodule contains Miocene Foraminifera such as *Nodogenerina avana* Cushman and Laiming (l. m. Miocene).

STATION 156

A nodule from this station contains a Mohnian fauna (l. u. Miocene).

STATION 158

Nodules have a mixed fauna including the Miocene form *Uvigerina hootsi* Rankin and a *Cassidulina* fauna of the Lomita Quarry type which is uppermost Pliocene to Recent.

STATION 159

Nodules contain *Valvulineria californica* Cushman, *Bolivina advena* Cushman, *Nonion costiferum* (Cushman), and others (l. m. Miocene).

STATION 162

The phosphorite is conglomeratic with a matrix having a well-preserved *Cassidulina* fauna of the Lomita Quarry type including *Cassidulina tortuosa* Cushman and Hughes, *C. subglobosa* var. *quadrata* Cushman, *C. limbata* Cushman and Hughes, *C. californica* Cushman and Hughes, *Bulimina ovula* d'Orbigny, and *Planulina ornata* d'Orbigny (uppermost Pliocene to Recent). An enclosed nodule contains a Miocene *Siphogenerina.*

STATION 183

Foraminifera present include *Baggina californica* Cushman, *B. robusta* Kleinpell, *Bulimina uvigerinaformis* Cushman and Kleinpell, *Bulimina ovula* d'Orbigny, *Cassidulina monicana* Cushman and Kleinpell, *Cassidulina modeloensis* Rankin, *Eponides healdi* Stewart and Stewart, *Nodogenerina advena* Cushman and Laiming, *Planulina ornata* d'Orbigny, *Virgulina californiensis* Cushman, *Uvigerina* cf. *senticosa* Cushman. Predominantly Mohnian (l. u. Miocene).

STATION 303

Foraminifera present include *Dentalina obliqua* (Linné) Woodring, Bramlette, and Kleinpell, *Nodogenerina advena* Cushman and Laiming, *Valvulineria* sp. Miocene or Pliocene type, *Bolivina* sp. Miocene or Pliocene type. Luisian or Relizian (m. Miocene).

STATION 9604

Bolivina girardensis Rankin is present. Relizian or Luisian (m. Miocene).

STATION 9610

A single *Siphogenerina* was found. (m. Miocene (?)).

Apparently the bulk of the material contains middle Miocene Foraminifera, but nodules from a few stations have upper Miocene Foraminifera. A Quaternary fauna which extends into uppermost Pliocene is present in some separate pieces of phosphorite and in layers superimposed upon a few nodules containing Miocene Foraminifera. The quantity of material containing this Quaternary fauna is relatively small as compared to the quantity containing the Miocene fauna. In one instance, a questionable Repetto (Pliocene) fauna was recognized.

Although both Recent and Miocene Foraminifera have been identified in some nodules, these faunas are confined to separate layers except in a very few specimens. While the identification is not certain, a number of nodules from Station 158 contain faunas composed of mixed Miocene, Pliocene, and Recent forms. A few other nodules contain a mixed fauna of Foraminifera characteristic of different faunal units of the Miocene.

The nodules containing Quaternary Foraminifera differ sufficiently in physical appearance from those containing Miocene tests to permit distinction on lithology alone. In general, they are characterized as follows: (1) They are light brown on fresh fracture; (2) they contain well-preserved Foraminifera which are usually unaltered or only partially replaced by francolite, and (3) they are conglomeratic, enclosing dark-brown to black pebbles of phosphorite having Miocene Foraminifera. Phosphorite containing Quaternary Foraminifera occurs as separate nodules and as layers superimposed upon some of the nodules having a Miocene fauna.

In contrast, most of the nodules having Miocene Foraminifera are characterized by the following: (1) They are dark brown to black, (2) they contain foraminiferal molds or poorly preserved tests completely replaced by francolite, and (3) they contain many layers, some of which are separated by manganese oxide.

Inasmuch as some nodules having a Miocene fauna are superimposed by a layer of different lithologic appearance which encloses a Quaternary fauna, it is apparent that there are at least two distinctly different times of phosphate deposition. Although phosphorite which contains the Quaternary Foraminifera must be of Quaternary age, those nodules having Miocene fauna were not necessarily deposited during the Miocene. Miocene Foraminifera may have been reworked into younger deposits

or phosphatic solutions may have infiltrated unconsolidated Miocene deposits. These considerations will be more fully discussed later in the paper.

DISTRIBUTION OF PHOSPHORITE

WORLD DISTRIBUTION

Continental distribution.—Marine deposits of nodular phosphorite, and bedded phosphorite into which they grade, are fairly common in the geological column, and range from Cambrian to Pleistocene. It has even been suggested (Dawson, 1876) that some of the rich apatite beds that are intimately associated with the pre-Cambrian Grenville marbles and gneisses near Ottawa, Canada, are highly metamorphosed representatives of phosphatic marine sediments. Among the best known phosphorite deposits are those of the Devonian of Tennessee, the Mississippian and Permian of the Rocky Mountains, the Cretaceous at Taplow in England, the Tertiary of Tunis and Algiers, and the Miocene and Pliocene of Florida. Some of these deposits extend over an area of thousands of square miles.

General oceanic distribution.—Since oceanographic expeditions have dredged phosphorite nodules in a considerable number of rocky portions of the ocean floor, it seems likely that phosphorite is relatively common on the ocean floor. The CHALLENGER Expedition (Murray and Renard, 1891) obtained the nodules on Agulhas Bank, south of the Cape of Good Hope; off the east coast of Japan; off the coast of Spain; off the east coast of Australia; off the coast of Chile; between the Falkland Islands; and off the mouth of the Plata River. More nodules were later dredged from Agulhas Bank by the German ships, GAZELLE and VALDIVA (cited by Collet, 1907), and by the Cape Government (Cayeux, 1934). The American ship BLAKE (Agassiz, 1888) dredged similar concretions off the east coast of the United States and in the Straits of Florida and later the ALBATROSS (Murray and Lee, 1909) obtained some nodules from the north Pacific. Published descriptions show that nodules from each of these localities agree closely in megascopic and microscopic character with those dredged off California. (*See* especially Cayeux, 1934.)

SUBMARINE DISTRIBUTION OFF SOUTHERN CALIFORNIA

General.—Plate 3 shows the location of stations where phosphorite was obtained off the southern California coast. Of 131 samples with rock, 43 contained phosphorite of a diameter greater than 5 millimeters. At some other stations partially phosphatized rock was found, and at others sand containing grains of phosphorite.

A near-shore zone, which is limited by the outer margin of the shelf, has a total of 49 rock hauls but only 8 of these contain phosphorite. It is notable that all but 2 of the latter were located off land areas where Miocene outcrops are dominant. The offshore zone beyond the shelf has a total of 82 rock hauls, of which 35 have phosphorite, a percentage of 43 compared with 16 per cent for the inshore zone. By subdividing the offshore zone into northern, central, and southern portions (Pl. 3), it was found that the southern portion yielded 6 rock hauls, none of which had phosphorite, whereas the central portion has phosphorite in 32 out of 65 rock hauls; and the northern portion in only 3 out of 11 hauls. These figures show that in the central portion 49 per cent of the hauls contain phosphorite compared to 33 per cent for the entire submarine area off southern California. Numerous Miocene outcrops have been found on the islands and the sea floor of this central portion.

Physiographic distribution.—Examination of Plate 3 shows that phosphorite is not only widespread, but also occurs in a considerable variety of topographic environments. It was obtained on the top and sides of banks, on steep escarpments which appear to be fault scarps, on walls of submarine canyons, and on the continental shelf. All of the localities where phosphorite is found are essentially nondepositional environments.

Dredgings brought up rock from nine canyons off the southern California coast. Phosphorite was obtained from the walls of three,—those off Redondo, Santa Monica, and Point Dume. Rock samples are available from 20 banks, 10 of which have yielded phosphorite. Since most of the banks which lack phosphorite have had only one dredging some of them may show the presence of phosphorite when more dredgings are available. San Juan Seamount, while not classed as a bank since it rises as a great mountain from the deep ocean floor, did not yield phosphorite. This seamount is thought to be a volcanic mass (Shepard and Emery, 1941, p. 28, 29).

The supposed fault scarps have not been very completely explored for phosphorite. However, the two counterposed scarps southeast of San Clemente Island, which form a V-shaped valley where they overlap (Shepard and Emery, 1941, Fig. 11) were dredged extensively and yielded phosphorite in 7 out of 17 hauls. Dredgings also brought up this material from other similar escarpments.

Rock has been dredged from eight localities on shelves including those around islands. Only the shelf off Santa Monica and that off San Diego yielded phosphorite. No phosphorite was found in the rocks from the shelf off Mission Beach (near San Diego), off Carlsbad, off Long Beach,

off San Nicolas Island, and on the small knob which rises above the shelf near the western end of the Santa Monica Mountains.

Depth and relation to shore.—There is little evidence of a lower depth limit for the material, nor any suggestion of a maximum concentration at any depth. The deepest dredging, which was taken at 10,320 feet on the continental slope, did not yield phosphorite, but it was found in a nearby sample at about 8400 feet. The CHALLENGER Expedition dredged some nodules from a depth of 11,400 feet at the base of the continental slope off the Cape of Good Hope. The shoalest sample of phosphorite was recovered from 240 feet on the Santa Monica Shelf. This haul (Station 303) contained an abundance of phosphorite. Rock dredged from 10 shoaler localities included no phosphorite. This does not prove that phosphorite is absent in shoal water, but it suggests this possibility. This is contrary to the hypothesis of Anderson (1895) that phosphorite predominates in littoral deposits.

Little phosphorite has been found close to shore, either on the continental shelf or on the insular shelves. On the other hand, almost half of the hauls from the offshore area contain phosphorite. The material is abundant in hauls made in the outer portions of the 150-mile wide continental borderland, but was not found in the few samples beyond the continental slope. Inasmuch as phosphorite has never been reported from the abyssal ocean basins, it may be that phosphorite, like glauconite, is formed only relatively close to shore.

Distribution in adjacent regions.—Although the data are few, phosphorite appears to occur in less abundance to the north and south of the southern California continental borderland. Twenty-three rock samples were collected by the E. W. SCRIPPS and the BLUEFIN to the north of this region, particularly in Monterey Bay. Phosphorite is present in 3 of these samples. In addition examination was made of samples collected by the Coast and Geodetic Survey along the west coast of North America. Over 100 of these sediment samples collected between Lats. 36° and 60° were studied. Although many of these contain rock fragments and are highly glauconitic, no phosphorite was found. In an expedition to the Gulf of California in 1940, 30 samples with rock were recovered of which 6 contained phosphorite.

OCCURRENCE ON LAND IN SOUTHERN CALIFORNIA

Although phosphorite is abundant on the sea floor, study of the literature, consultation with other geologists, and field work have shown that it is relatively uncommon on the mainland and apparently absent on the islands. Miocene and to a lesser extent Pliocene and Pleistocene rocks

on the mainland contain a small amount of phosphorite. Drilling in the Playa del Rey and the Torrence oil fields on the west end of the Los Angeles Basin encountered 150 feet of upper Miocene black shale (Hoots, *et al.*, 1935) which has a spotted and a nodular nature due to the erratic occurrence of tan and gray phosphorite nodules from 1 to 8 centimeters in diameter. Some of these are porous and a few have an indistinct oölitic structure. The nodules contain about 70 per cent tri-calcium phosphate and are associated with thin discontinuous tan-colored bands of the same composition; however, they are distinctly different in size, shape, and structure from those dredged offshore.

A dark-colored bituminous nodular shale 200 feet thick, also upper Miocene, is exposed half a mile south of Capitan Point, north of Ventura. In many ways this deposit is similar to that of the oil fields. Inasmuch as the phosphorite nodules occur along bedding planes and not along joints and since they are in a fine-grained deposit, they were apparently formed syngenetically and in place.

One of the most interesting deposits of phosphorite found on land in California is at the Lomita Quarry on the north side of the Palos Verdes Hills. Pieces of phosphorite as long as 25 centimeters occur in a glauconitic marl and also in association with a band of subangular to subround rocks at a slightly higher horizon. In this band the nodules predominate over other types of rock. Although the enclosing beds are lower Pleistocene (or uppermost Pliocene) (Galloway and Wissler, 1927), M. L. Natland and H. L. Driver (personal communications) have identified Miocene Foraminifera in the phosphorite. Consequently the nodules, or at least the Foraminifera, have been reworked out of Miocene formations. In contrast to the phosphorite from the Miocene nodular shale, the material from the Lomita Quarry resembles that dredged from the ocean floor in color and in the presence of oölites, banding, glauconite, and Foraminifera.

Another occurrence of phosphorite is in a near-by road cut a quarter of a mile south of the Lomita Sand and Gravel Company. These nodules are in a well-sorted basal gravel bed unconformably overlying tilted Miocene beds. About one third of the pebbles in one zone are phosphorite.

ORIGIN OF THE PHOSPHORITE

PREVIOUS VIEWS

Many writers have attempted to explain the origin of phosphorite and the physical-chemical conditions which caused the precipitation of phosphate. Blackwelder (1916) and Mansfield (1927, p. 365) suggested that this precipitation may take place in sea water low in oxygen. These

writers have pointed out that an oxygen deficiency exists in submarine basins where stagnant conditions are reflected in the high content of organic matter in the sediments. This stagnant water inhibits the growth of organisms and may allow phosphate to become sufficiently concentrated to produce chemical reactions that are not ordinarily operative. The ion thus becomes fixed in the form of nodules or continuous beds of phosphorite. Although this may be the correct explanation for many phosphorites and for phosphatic shales, it probably does not account for the phosphatic nodules off California since the latter appear to have formed in association with coarse-grained glauconitic sands on topographic highs. Furthermore, it has been shown (Trask, 1932, p. 120; Revelle and Shepard, 1939, p. 262) that the sediments from the topographic highs off California where the phosphorite is found are relatively low in organic matter. In addition, oceanographic data show that these areas are among the most oxidizing environments of the sea floor (Revelle, personal communication).

In order to account for the occurrence and distribution of the phosphorite nodules dredged by the CHALLENGER Expedition, Murray and Renard (1891, p. 396-399) postulated a catastrophic destruction of life resulting from the mixing of a cold water mass with a warm water mass as might occasionally occur along the border of a great ocean current. According to Murray and Renard, organic matter would be deposited in such abundance that special chemical conditions might operate so that phosphate would not be returned to the sea water, but rather would be precipitated as nodules. Later, Murray (1898, p. 113) cited a remarkable case in which tile-fish were killed by the hundreds of millions along the Atlantic Coast of the United States in 1883, presumably due to the shifting of oceanic currents. However, Blackwelder (1916, p. 290) has pointed out that dead fish float and Mansfield (1927), in attempting to apply this theory to the Phosphoria formation, states that the relative rarity of animal remains such as fish teeth is incompatible with catastrophic killing. The same objections apply to such an origin for the deposits off the California coast.

Pardee (1917) has suggested that the deposition of the Phosphoria formation was due to the same cool climatic conditions that produced the glaciers known to have existed during the Permian period. Carbon dioxide, supplied not only by the atmosphere but also by decaying organic matter on the sea bottom, is retained most abundantly by waters of low temperature. In such water calcareous remains would tend to be dissolved and the accumulation of lime deposits would be hindered; consequently, the phosphatic substances would accumulate in relatively pure

form. However, it is apparent that a glacial climate is not essential for the deposition of phosphorite because similar deposits range from Cambrian to Recent and widespread glacial conditions were infrequent during this interval. The possible Pleistocene age of the deposit off California might be thought to support this theory of Pardee, but most of the nodules were dredged from depths greater than 1000 feet, and some of them also contain deep-water Foraminifera indicating that the nodules were formed at depths similar to those in which they are now found. Sea water this far below the surface has a uniform temperature of a few degrees above zero centigrade; consequently, it would be possible for this water to become only slightly colder. Such a small change probably would not exert an important influence upon the chemical equilibrium of the sea water.

Mansfield (1940b) has suggested that fluorine may be responsible for the deposition of phosphorite. He pointed out that most phosphorites are composed of fluorapatite and were probably originally deposited as this material.

Since fluorine in the form of HF is a common constituent of volcanic gases, Mansfield maintains that times of intense volcanism are favorable to the deposition of phosphorite. The California phosphatic shales are Miocene,—a time of widespread volcanism in this region (Reed, 1933, p. 203).

CHEMISTRY OF PHOSPHATE IN THE OCEAN

General.—The mineral apatite, a minor primary constituent of practically all igneous rocks, is regarded as the principal original source of the sedimentary phosphates (Blackwelder, 1916, p. 286). The phosphorus liberated in the decomposition of rocks on land is ultimately carried in solution to the sea where plants and animals extract small quantities for use in their body structures. Phosphorus is an essential constituent of protoplasm and an important constituent of bones, teeth, and many shells. After the death of an organism, its remains are devoured by scavengers and decomposed by bacteria, with the result that the phosphorus is returned to the ocean water. From time to time, phosphorus escapes this apparently endless cycle and becomes fixed in the sedimentary deposits. An understanding of this fixation process requires a consideration of the chemistry of phosphate in sea water.

Solubility data for $Ca_3(PO_4)_2$.—It was felt that an investigation of the chemistry of phosphates in sea water might lead to a clearer understanding of the methods by which phosphorite and phosphatic shales are formed. So far as the writers could determine, no data are available in the literature regarding the solubility of phosphorite nor of its components, collophane and francolite. However, some experimentally deter-

mined constants for $Ca_3(PO_4)_2$ have been published by Sendroy and Hastings (1927). They found the following equation to hold in a salt solution at 38° C. where μ is the ionic strength of the solution.

$$PK'_{S.P.\ Ca_3(PO_4)_2} = 30.95 - \frac{17.40\mu^{\frac{1}{2}}}{1+1.14\mu^{\frac{1}{2}}} \qquad (1)$$

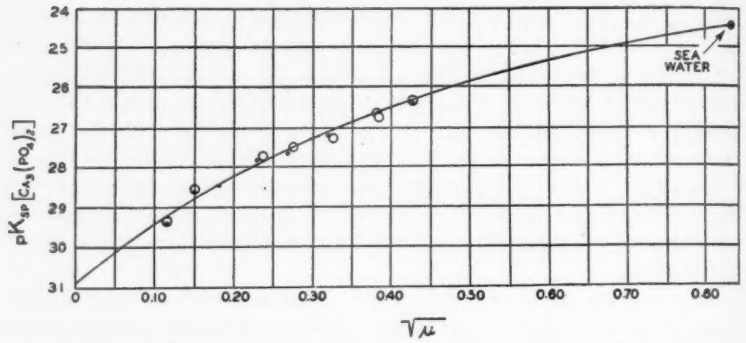


FIGURE 1.—*Relationship between ionic strength* (μ) *and solubility of* $Ca_3(PO_4)_2$
Showing extrapolation of Sendroy and Hastings data (circles, dots, and crosses) to sea water.

The ionic strength of sea water may be determined from the equations given by Lyman and Fleming (1940, p. 141), namely:

$$\mu \text{ for sea water} = 0.00147 + 0.03592\ (\text{Cl}‰) + 0.000068\ (\text{Cl}‰)^2 \qquad (2)$$

where (Cl‰) is the chlorinity in grams per kilogram. The chlorinity of the sea water averages about 19‰ but may be only 18‰ or 17‰ at relatively slight depths (Thompson and Robinson, 1932, p. 112). Accordingly substituting from (2) into (1) above,[3] when the chlorinity is approximately 19‰:

$$[Ca^{++}]^3 \times [PO_4^{\equiv}]^2 = K'_{S.P.\ Ca_3(PO_4)_2} = 3.73\,(10)^{-25} \qquad (3)$$

Substituting these values in (3) and transposing:

$$[PO_4^{\equiv}] = \sqrt{\frac{3.73\ (10)^{-25}}{(0.010)^3}} = 0.61\,(10)^{-9} \text{ mols per liter} \qquad (4)$$

This value represents the amount of $[PO_4^{\equiv}]$, which theoretically can be present in sea water.

From oceanographic work, total phosphate is known to increase from about zero at the surface to over 200 milligrams per cubic meter at depths

[3] The ionic strength of sea water is greater than that of the salt solution used in the experiments of Sendroy and Hastings. However, the calculated sea water values for $\mu^{\frac{1}{2}}$ of 0.84; and for $K'_{S.P.\ Ca_3(PO_4)_2}$ of 3.73 (10)$^{-25}$ probably represents a safe extrapolation of their data (Fig. 1).

of several thousand meters in the open sea. Values of phosphate given in oceanographic reports are simply the sum of the PO_4 radicals, no attempt being made to determine the concentration of the dissociated ions. However, the dissociation constants of phosphoric acid used by Sendoy and Hastings (p. 786-790, 801) may possibly be extrapolated for sea water. For a solution in which $\mu = 0.71$, these constants are:

$$PK'_1 = 1.69$$
$$PK'_2 = 6.10$$
$$PK'_3 = 10.77$$

Thus, the corresponding values of K′:

$$\frac{(H^+) \times [H_2PO_4^-]}{[H_3PO_4]} = 2.04\ (10)^{-2} \tag{5}$$

$$\frac{(H^+) \times [HPO_4^{=}]}{[H_2PO_4^-]} = 7.95\ (10)^{-7} \tag{6}$$

$$\frac{(H^+) \times [PO_4^{\equiv}]}{[HPO_4^{=}]} = 1.70\ (10)^{-11} \tag{7}$$

In sea water having a total phosphate content of 100 milligrams per cubic meter, or 1.05 $(10)^{-6}$ mols per liter:

$$[H_3PO_4] + [H_2PO_4^-] + [HPO_4^{=}] + [PO_4^{\equiv}] = 1.05\ (10)^{-6} \tag{8}$$

Thus, if values of (H^+) are substituted in (5), (6), and (7) we have four equations with four unknowns, namely, (5), (6), (7), and (8) from which $[PO_4^{\equiv}]$ may be determined. From these equations and at a pH of 8, $[PO_4^{\equiv}]$ is 1.76 $(10)^{-9}$ mols/l. Dividing this value by the theoretical saturation concentration (4), we find that such a solution would be 289% saturated.

Sources of error.—In the derivation of the solubility of $Ca_3(PO_4)_2$ use was made of the solubility product determined at 38° C., and the best available ionic dissociation constants, some of which were also determined at 38° C. However, sea water always has a much lower temperature, and at depths of a few hundred meters in the open sea is less than 10° C.; accordingly these constants may be somewhat incorrect.

The solubility of $Ca_3(PO_4)_2$ is undoubtedly influenced to some extent by pressure (Gibson, 1938), and by temperature, but relatively little is known of the effect of these factors. Another possible source of error exists in the unknown effect of certain other ions which may make $Ca_3(PO_4)_2$ more or less insoluble. One such ion is F^-, which, according to Mansfield (1940b, p. 867) may be responsible for the deposition of phosphate in a relatively insoluble form. Adler, *et al.* (1938) point out that $Ca_3(PO_4)_2$ is an effective adsorbent of F in ground waters.

Further error results from the use of chemical data based on $Ca_3(PO_4)_2$, whereas phosphorite is a tri-calcium phosphate containing water, carbonate, and fluorine. In addition, the law of Debye and Hückel (which states that the activity of an ion in salt solution is dependent on the ionic strength of that solution) is correct only as a first approximation.

Application of the solubility data.—Although these calculations concerning the degree of saturation of sea water with tri-calcium phosphate can be only approximate, significant conclusions are suggested. For example, it appears probable that sea water deeper than a few hundred meters is essentially saturated with tri-calcium phosphate. If the ocean is saturated, it follows that an amount equal to all the phosphate annually carried to the sea by rivers must be deposited as phosphate minerals or as organic matter. Also such saturation may allow the existence of a colloidal phase of tri-calcium phosphate in equilibrium with the dissolved phase. It has been pointed out above that phosphorite is composed of colloidal particles, and this is at least suggestive that precipitation took place from a colloidal suspension.

Owing to the dependence of the phosphate saturation point on pH, and the wide variation in phosphate content in the sea, slight changes in physical-chemical or biological conditions may bring about supersaturation or undersaturation. Changes brought about by the activity of benthonic organisms may cause solution or precipitation of phosphorites. Thus the calcium phosphate balance in the sea is similar to that of calcium carbonate.

DEPOSITION ESSENTIALLY *IN SITU*

Nodules of phosphorite may form *in situ* or may have been transported to their present positions. The following considerations suggest that a large portion of the phosphorite off the California coast was formed essentially in place:

(1) The mammillary surfaces formed by growth layers on many nodules are present on only one side, and these same sides have attached organisms and coatings of manganese oxide. This association indicates that at least many of these particular nodules have grown in place and have not been subsequently disturbed.

(2) The very large size of many of the nodules and their lack of abrasion rounding suggest absence of transportation and even of disturbance by waves after the nodules were formed.

(3) The discovery of sediment within the nodules which closely resembles the surrounding sediment suggests that the nodules have formed in place. For the same reason Murray and Renard (1891, p. 397) came

to the conclusion that the nodules on the sea floor off the Cape of Good Hope were formed *in situ*.

(4) Attention has been called previously (Goldman, 1922) to nondepositional environments in the stratigraphic column as favorable for the accumulation of nodular phosphorite. The phosphorite is found in just such environments off southern California.

Additional considerations which oppose long distance transportation include: (1) There should be some remnant of the source beds on land, and (2) no method is known whereby the largest nodules could have been transported any appreciable distance from their source beds.

ORIGIN OF NODULES

Phosphorite nodules and closely related beds of marine phosphorite have been explained by direct deposition from solution or by replacement of other rocks. Murray and Renard (1891, p. 399) suggested that colloidal calcium phosphate supplied from decaying organisms is precipitated about some nucleus and then grows by attraction of like particles. These writers also pointed out that ammonium phosphate liberated by organisms may replace calcium carbonate. In support of this idea, they cited the experiment of Irvine and Anderson (1891, p. 52-54) in which the immersion of a porous variety of coral in a solution of ammonium phosphate for 6 months resulted in a replacement of 60 per cent of the calcium carbonate by calcium phosphate. Blackwelder (1916) has also suggested that ammonical solutions, derived from the decay of marine organisms, dissolve phosphatic material from bones, teeth, and shells. This phosphate may then locally replace carbonates and enrich the calcium phosphate in bones and phosphatic shells, although Blackwelder thought it would be precipitated directly to a greater extent, forming nodules and bedded phosphorite. Replacement of rock by phosphatic solutions has also been reported in localities where the rock was overlain by guano (Clark, 1924, p. 525).

The following considerations appear to indicate that the nodules off California are primary, that is, the result of direct precipitation:

(1) Practically all of the nodules are composed of irregular nonconcentric layers of brown collophane separated by a thin film of darker shade. These films have all the characteristics of the coatings found on the surface of the nodules, indicating that they are former surface coatings and showing that the nodules were formed by discontinuous accretion. The lapse of time between these periods of accretion is commonly shown by a coating of manganese oxide and by the presence of a band of glauconite grains above the dark collophane film.

(2) Tarr and Twenhofel (Twenhofel, 1932, p. 533-535) have shown that colloidal particles of silica because of their size probably lack the ability to penetrate dense limestone and consequently cannot replace them. Tarr's observations suggest that fossils in chert and elsewhere are never replaced by chert but rather by quartz. Similarly, the foraminiferal tests in phosphorite nodules are not composed of microcrystalline colloidal collophane but rather of francolite apparently formed by ionic replacement. The nodules, however, are largely composed of collophane, the colloidal particles of which could not have penetrated and replaced dense limestone. Some phosphorite was found coating massive limestone, but this coating is separated from the limestone by a sharp contact beneath which the limestone is essentially unaltered.

(3) The nodules have distinctive textural features unlike those which would be expected if a limestone or a related sedimentary rock had been replaced. In this connection, the enclosure within the nodule of bones, glauconite, coarse clastic mineral grains and rock fragments is significant because their abundance is greater than is usual in calcareous rocks.

Although the majority of the nodules are apparently of primary origin, most of them enclose a considerable amount of replaced material that was originally calcareous. Foraminiferal tests replaced by francolite are extremely common and form an important part of some of the nodules. In all thin sections, scattered anisotropic particles of francolite are present. Probably these are replacements of finely comminuted calcium carbonate, like that found within the associated sediments.

Much of the calcareous sediment associated with the nodules gives a strong chemical test for phosphate indicating that some phosphatization has taken place. A few pieces of homogeneous phosphorite, which are not of nodular form, may be replaced limestone. The dredging of slightly phosphatic limestones which are similar in appearance to these homogeneous pieces of phosphorite lends some support to this suggestion.

ORIGIN OF OÖLITES

Noncalcareous oölites in many deposits have been explained as originally formed of calcium carbonate and later replaced by some other substance. Mansfield (1918, p. 591) at first offered such an explanation for the oölites of the Phosphoria formation. He suggested that the calcium carbonate of the original oölites was replaced by calcium phosphate through the action of phosphatic solutions produced from bacterial decay of marine organisms. He later decided (1927, p. 364) that these phosphatic oölites were probably formed directly from phosphatic colloidal suspensions on the sea bottom through biochemical and physical agencies. As indicated above, most of the sea floor nodules and the

associated sediment contain ovoid oölites. The concentric layering of the oölites indicates a process of accretion, but their oval shape is puzzling in view of the usual spherical shape of oölites.

In some samples oval pellets are found which are composed of finely comminuted limy debris with neither nuclei nor concentric layering. Their appearance suggests that they may have been formed and molded into their present oval shape by being passed through the intestines of marine mud-eating animals. The fine fractions of sediment in which these pellets are found are composed almost entirely of calcium carbonate. Pellets in all stages of phosphatization are present, ranging from those which are composed of unaltered calcium carbonate to those which have been completely phosphatized both by replacement and by filling of interstices.

TIME OF PHOSPHORITE DEPOSITION

GENERAL CONSIDERATIONS

Regardless of the method of formation, a separate problem is involved in determining the time and manner in which the nodules were collected into the deposits in which they were found. Hypotheses must explain two facts of particular significance:

(1) Phosphorite nodules have a wide areal distribution, being obtained at 43 stations off southern California. This material was recovered in greater quantities and from more stations than any other type of rock. Several large hauls consisted almost entirely of these nodules. The wide surface distribution is in marked contrast to the known distribution of phosphorite on land in California or elsewhere. In a region of outcropping beds of phosphorite, such as that of the Permian Phosphoria formation of the Rocky Mountains, the phosphorite forms only a small percentage of all the rocks exposed at the surface. If such an area were submerged intact, there would be little chance that dredging would recover large amounts of phosphorite such as were found off southern California. This implies that the phosphorite was obtained from a surface covering rather than from the outcropping edge of a buried formation; in other words the nodules were dredged from a single horizon rather than from a stratigraphic section.

(2) The nodules contain a predominance of Miocene Foraminifera. Those from 16 stations enclose identifiable Foraminifera, which are dominantly middle and upper Miocene. In nodules from at least one station, a mixed fauna of Miocene and Pliocene Foraminifera is present. In several instances, separate nodules and phosphatic cement of conglomerates contain Quaternary Foraminifera.

POSSIBLE MIOCENE AGE OF PHOSPHORITE

General.—The data given above indicate that at least a small portion of the phosphorite must be Quaternary; however, there remains the problem of whether the bulk of the phosphorite which covers so much of the submerged banks, slopes, and shelves is Miocene as indicated by the foraminiferal content or whether it can be explained by deposition of phosphatic material at a more recent time. There are at least two ways in which the phosphorite could have been formed during the Miocene, and still be abundant: (1) The phosphorite might have formed during the Miocene on submarine highs which have been neither eroded nor covered with sediment since that time; (2) the nodules on the present sea floor may have undergone residual concentration from Miocene rocks.

Both of these explanations appear to be favored by the following points:

(1) Miocene Foraminifera are predominant within the nodules.

(2) The phosphorite containing Miocene Foraminifera is different from that enclosing Quaternary Foraminifera.

(3) Phosphorites on land in southern California are more abundant in rocks of Miocene than in those of other ages but are entirely different from those on the present ocean floor.

(4) Almost all of the phosphorite is associated with nonphosphatic Miocene rocks, whereas the 12 dredgings consisting entirely of sedimentary rocks of other ages lack phosphorite.

(5) Phosphorite is more abundant in dredgings on the shelf and in the canyons adjacent to land where Miocene rocks outcrop.

(6) The transported phosphorite nodules of the Lomita Quarry also contain Miocene Foraminifera, but are found in a lower Pleistocene (or upper Pliocene) formation, and hence were originally pre-Pleistocene.

(7) In the Gulf of California, phosphorite similar to that found off California contains Miocene Foraminifera.

Formed in nondepositional areas persisting since Miocene.—Submerged, nondepositional highs may have developed in the continental borderland area in middle Miocene since this is said to have been a time of widespread faulting and basin development in California (Reed, 1933, p. 286); and they may have remained above the adjacent basins and yet below the surface of the ocean since that time. If this were true, the nondepositional nature of such topographic highs (Shepard, 1941) should have prevented the covering of these surfaces with clastic sediments and phosphorite deposited in the Miocene might have remained uneroded. This hypothesis reconciles the data suggesting *in situ* deposition with the paleontological evidence. Supporting both the suggestion that the banks and

the basins were initiated by basin development in the Miocene and that they have remained essentially nondepositional environments since that time, is the fact that nearly all the nonphosphatic sedimentary rocks dredged from these banks are Miocene. Practically no Pliocene material has been obtained.

However, there are many objections to this hypothesis. The continental borderland off southern California is known to be unstable as shown by the elevated terraces of probable Quaternary age on San Clemente, San Nicolas, and the Santa Barbara Islands. Relatively recent submarine fault scarps are found, as for example in the vicinity of San Clemente Island (Shepard and Emery, 1941, p. 21-25). The presence of phosphorite with Miocene Foraminifera on some of these fault scarps forms a serious objection to the hypothesis under consideration since in all probability the surfaces do not date back to the Miocene. There are insurmountable difficulties in attributing continental shelves such as that of Santa Monica Bay to wave erosion during the Miocene and still more serious difficulties in dating as Miocene the canyons which cut this Santa Monica Bay shelf. Yet both the shelf and the canyon walls have considerable quantities of phosphorite, some of which contains Miocene Foraminifera.

The existence of the outlying banks since the Miocene might have led to the deposition of Pliocene phosphorite on these surfaces, but there is little evidence that nodules of this age are present. It is incredible that extensive deposits of phosphorite formed on banks during the Miocene could have remained uncovered and free from erosion in all the intervening time. Evidence of the elevation of the banks and shelves comes from the presence of canyons with fluviatile characteristics cut into them and from the apparent wave eroded surfaces and the rounded cobbles found on many of these highs.

The Miocene cherts, limestones, and mudstones differ greatly from the glauconitic sediments found on the banks and were presumably deposited in broad embayments. Therefore, the banks must have been initiated, that is, elevated above their surroundings, since deposition of these Miocene rocks. If this were true, the phosphorite should contain more recent faunas than the other rocks obtained in the same dredging, but the available data show that the Foraminifera of the phosphorite are in general older than those of the associated Miocene rocks.

Formed by residual concentration of Miocene phosphorite nodules.—Another hypothesis compatible with a Miocene age of the bulk of the phosphorite is that the nodules were concentrated on submarine surfaces by differential marine erosion of soft Miocene beds containing many phos-

phorite nodules. A similar origin is postulated for the Pliocene phosphorites of Florida (Mansfield, 1940a, p. 412, 414, 415). This process would account for the abundance of the phosphorite even where it is probable that the surfaces have been formed in relatively recent time as in the case of the San Clemente fault slope, the Santa Monica shelf, and the walls of submarine canyons.

There are a number of serious objections to this hypothesis. The great majority of the Miocene strata on land contain relatively few phosphorite nodules either in surface outcrops or in well cores. While nodules have been found in extensive drilling operations in Miocene rocks none like those dredged on the sea floor have been reported. The only nodules of phosphorite like those of the sea floor were found reworked into beds of Pleistocene or uppermost Pliocene age in Lomita Quarry and vicinity. Certainly the marine weathering and erosion of a normal Miocene outcrop of the Los Angeles area would yield very little phosphorite. Examination of pebbles on elevated marine terraces of the Palos Verdes Hills failed to reveal any phosphorite despite the fact that these terraces were cut into Miocene formations. Also, phosphorite was not found in the Miocene of San Clemente and Catalina Islands.

A considerable number of the nodules appear to have been deposited *in situ* and to have remained undisturbed since their deposition. If these nodules are residual concentrations from Miocene deposits, their positions would have been disturbed. Furthermore, the impurities in the nodules are similar to the sediment found on the surfaces of slow deposition where the nodules were discovered. No known Miocene sediments are sufficiently similar to those of the present banks for this resemblance to hold.

Therefore, this hypothesis requires the formation of a different type of Miocene deposit in the area constituting the present sea floor from that which accumulated in the area of the present continent and islands. However, the nonphosphatic rocks containing Miocene fossils dredged from the continental borderland give no indication of an environment different from that suggested by the continental and insular outcrops of Miocene age.

POSSIBLE QUATERNARY (OR PLIOCENE) AGE OF PHOSPHORITE

General.—A few nodules of phosphorite and the matrix of the phosphorite conglomerates contain Quaternary Foraminifera. On the other hand, most of the nodules have Miocene Foraminifera. The two following explanations could account for the Miocene Foraminifera despite a Quaternary age for the nodules.

(1) The phosphorite deposits may have been formed by the infiltration of phosphatic solutions into porous Miocene formations with the replacement of some of the Miocene material.

(2) The phosphorite deposits may have been formed by enclosure of reworked Miocene Foraminifera in nodules formed during the Quaternary.

Because dredging of the continental borderland off southern California has shown the preponderant presence of Miocene rocks, both replaced and reworked material should contain some Miocene Foraminifera. Points favoring a Quaternary or at least a post-Miocene origin are as follows: (1) The predominance of phosphorite over other types of rock dredged from the sea floor off this coast, (2) correspondence of the impurities in the phosphorite nodules with the associated sediments, (3) virtual limitation of the nodules to the present sea floor, suggesting a relation to the present seas.

Formed by infiltration and replacement of Miocene limestone.—Since loosely consolidated porous Miocene formations have been exposed on the sea floor over extensive areas during the Quaternary, phosphate solutions may have infiltered these deposits, completely filling the pore space. Precipitation of phosphorite out of these solutions would have produced a rock with a high percentage of phosphorite. More complete phosphatization would have resulted in replacement by francolite of the Foraminifera and other limy portions of the Miocene rock. Some continental phosphatic deposits have been explained in this way as well as some of the material from Agulhas Bank (Cayeux, 1934, p. 130-132).

This infiltration and replacement may have taken place as the result of contact with sea water locally supersaturated with phosphate. Also, there seems to be considerable evidence that the banks off southern California have been elevated to or above sea level (Shepard and Emery, 1941, p. 16). If the banks were exposed as islands they may have become sites of guano deposition. Since phosphatization of rock underlying guano is known to occur, phosphatic solutions from the guano may have produced some infiltration and replacement.

However, the accretionary layers of most of the phosphorite nodules are highly inconsistent with a replacement origin. Also, available samples of rock, which had been replaced by phosphatic solutions, are quite unlike the material found on the sea floor off California. Therefore, this explanation can apply at most to a small portion of the phosphorite.

Formed in reworked Miocene deposits.—There are numerous Pliocene and Quaternary beds which contain a considerable percentage of Miocene Foraminifera (M. L. Natland, personal communication). Also, at many localities on the sea floor, sands have been encountered which are charac-

terized by an abundance of Miocene Foraminifera. If Miocene Foraminifera were abundant on the sea floor, phosphorite of Quaternary age could incorporate these Foraminifera into nodules. Nodules would also contain glauconite and fragments of underlying formations found on these surfaces of nondeposition. All these impurities are present. It is possible that associated with the Miocene Foraminifera are a few Quaternary forms which have been overlooked because of their scarcity. A mixed assemblage of Miocene, Pliocene, and Quaternary Foraminifera was found in phosphorite from one locality. Small nodules of phosphorite such as the Miocene does contain may possibly have served as nuclei for the growth of the larger nodules. Therefore, the previous existence of phosphorite in the Miocene might have favored further deposition on surfaces containing eroded material of that age.

An objection to this mode of origin is that mixed faunas of Miocene, Pliocene, and Quaternary Foraminifera have not been more commonly identified in the nodules. Furthermore, the nodules frequently contain nests of a single species of Foraminifera, such as a Miocene *Siphogenerina*. Such a nest is difficult to explain if the foraminiferal tests were reworked. In a reworked formation above an unconformity ordinarily there would be an admixture of contemporaneous and older forms but the latter may greatly predominate. The surfaces under discussion are unconformities in the making. Also changes in sea level such as accompanied the Pleistocene or changes in the chemistry of the sea water may have produced conditions which led to the exclusion of Recent Foraminifera from the sediment in which the nodules were formed.

SUMMARY AND CONCLUSIONS

Nodular phosphorite was dredged in great abundance from the sea floor off southern California, some hauls containing hundreds of nodules. These nodules have protuberances and cavities with smooth, glazed, brown surfaces. They are composed of irregular layers of collophane and usually enclose scattered phosphatic oölites, glauconite, and foraminiferal tests. Clastic mineral grains, rock fragments, and phosphatized bones are also commonly present. The nodules agree closely with published descriptions of those obtained from the ocean floor in other parts of the world.

Topographic environment is a most important factor in determining the distribution of the phosphorite. The material has been found in great abundance on the offshore banks and steep slopes and has also been recovered from five submarine canyons and from one area on the continental shelf. The nondepositional nature of the surface on which the nodules are found is indicated by the presence of rock bottom partly covered with a thin patchy veneer of glauconitic foraminiferal sand.

Most of the phosphorite appears to have been formed by direct precipitation from ocean water although it encloses foraminiferal tests and other limy materials which have been replaced by francolite. Layering of the phosphorite indicates discontinuous accretion of collophane. In many cases sufficient time elapsed between periods of accretion to allow the deposition of a coating of manganese oxide. Facts indicative of origin in place include: the great abundance, wide areal distribution, lack of abrasion, large size, poor sorting, and similarity between the enclosed material and the surrounding sediment.

Calculations indicate that ocean water is essentially saturated with tri-calcium phosphate. Hence, slight changes in the physical-chemical environment might be expected to cause the direct precipitation of tri-calcium phosphate on the sea floor. However, the material that is actually precipitated is not tri-calcium phosphate but is a compound or mixture of compounds having the apatite structure.

Foraminifera in a small part of the phosphorite are of Quaternary (or upper Pliocene) age, establishing deposition in relatively recent time. Most of the nodules, however, contain Miocene Foraminifera or resemble other nodules which contain Miocene Foraminifera. While this faunal content may indicate Miocene deposition of the phosphorite the following points suggest a more recent origin:

(1) The deposits are far too extensive to be erosion remnants of a special type of deposition during the Miocene.

(2) The impurities in the nodules indicate deposition in areas where clastic sediments were not accumulating, a condition such as exists now where the nodules are found. The borderland is too unstable to account for the existence of the present surfaces without erosion or deposition since the Miocene. Furthermore, the Miocene formations from the sea-bottom in this area indicate that ordinary deposition was progressing both during and after the time when the faunas found in the phosphorite were living.

(3) The possibility that the nodules are residual concentrates of nodules from the Miocene formations exposed on the banks is opposed by the failure to find similar concretions in the Miocene on land and by the evidence of lack of disturbance of many nodules since deposition.

These considerations lead to the tentative conclusion that the nodules were deposited on the present banks, shelves, and canyon walls during the Quaternary and that previously an abundance of Miocene Foraminifera had been eroded or weathered out of the Miocene formations and concentrated on the surfaces on which phosphorite deposition took place.

REFERENCES CITED

Adler, H., Klein, G., Lindsay, P. K. (1938) *The removal of fluorides from potable water by tricalcium phosphate,* Ind. Eng. Chem., vol. 30, p. 163-165.

Agassiz, A. (1888) *Three cruises of the Blake,* vol. 1, Mus. Comp. Zool., Bull., Harvard College, vol. 14, p. 281-282.

Anderson, J. G. (1895) *Über Cambrische und Silurische phosphoritführende gesteine aus Schweden,* Bull. Geol. Ind. Univ. Upsala, 2, p. 178-200, 220-229.

Blackwelder, Eliot (1916) *The geologic role of phosphorus,* Am. Jour. Sci., 4th ser., vol. 42, p. 285-298.

Cayeux, L. (1934) *The phosphate nodules of Agulhas Bank,* S. African Mus., Ann., vol. 31, p. 105-136 (trans. by S. H. Haughton).

Clark, F. W. (1924) *Data of geochemistry,* U. S. Geol. Survey, Bull. 770, p. 523-534.

Collet, L. W. (1907) *Les dépôts marines,* p. 194-213, Paris.

Dawson, J. W. (1876) *Note on the phosphate of the Laurentian and Cambrian rocks of Canada,* Geol. Soc. London, Jour., vol. 32, p. 285-291.

Emery, K. O., and Shepard, F. P., *Lithology of the sea floor off southern California* (in manuscript).

Galloway, J. J., and Wissler, S. G. (1927) *Pleistocene Foraminifera from the Lomita Quarry, Palos Verdes Hills, California,* Jour. Paleont., vol. 1, p. 35-87.

Gibson, R. E. (1938) *On the effect of pressure on the solubility of solids in liquids,* Am. Jour. Sci., 5th ser., vol. 35A, p. 65.

Goldman, M. I. (1922) *Basal glauconite and phosphate beds,* Science, vol. 56, no. 1141, p. 171-173.

Hendricks, S. B., Hill, W. L., Jacob, K. D., and Jefferson, M. E. (1931) *Structural characteristics of apatite-like substances and the composition of phosphate rock and bone as determined from microscopical and x-ray diffraction examinations,* Ind. Eng. Chem., vol. 23, p. 1413-1418.

Hoots, H. W., Blount, A. L., and Jones, P. H. (1935) *Marine oil shale, source of oil in Playa del Rey field, California,* Am. Assoc. Petrol. Geol., Bull. 19, pt. 1, p. 180.

Irvine, Robert, and Anderson, W. S. (1891) *On the action of metallic (and other) salts on carbonate of lime,* Royal Soc. Edinburgh, Pr., vol. 17, p. 42-52.

Kleinpell, R. M. (1938) *Miocene stratigraphy of California,* Am. Assoc. Petrol. Geol., Tulsa.

Lyman, J., and Fleming, R. A. (1940) *Composition of sea water,* Jour. Marine Res., vol. 3, p. 134-146.

Mansfield, G. R. (1918) *Origin of the western phosphates of the United States,* Am. Jour. Sci., 4th ser., vol. 46, p. 591-598.

——— (1927) *Geography, geology, and mineral resources of part of southeastern Idaho,* U. S. Geol. Survey, Prof. Paper 152, p. 361-367.

——— (1940a) *Phosphate deposits of the United States,* Econ. Geol., vol. 35, no. 3, p. 405-429.

——— (1940b) *The role of fluorine in phosphate deposition,* Am. Jour. Sci., vol. 238, p. 863-879.

Murray, J. (1898) *On the annual range of temperature in the surface waters of the ocean, and its relation to other oceanographical phenomena,* Geogr. Jour., vol. 12, p. 113.

———, and Lee, G. V. (1909) *Depth and marine deposits of the Pacific,* Mus. Comp. Zool., Mem., Harvard College, vol. 38, no. 1, p. 35.

Murray, J., and Reynard, A. F. (1891) *Scientific results, H. M. S. Challenger, Deep sea deposits,* p. 391-400.

Pardee, J. T. (1917) *The Garrison and Phillipsburg phosphate fields,* U. S. Geol. Survey, Bull. 640, p. 224-228.

Reed, R. D. (1933) *Geology of California,* Am. Assoc. Petrol. Geol., Tulsa, Okla.

Revelle, R. R., and Shepard, F. P. (1939) *Sediments off the California coast,* Symposium on Recent Marine Sediments, Am. Assoc. Petrol. Geol., p. 245-282.

Sendroy, J., and Hastings, A. B. (1927) *Studies of the solubility of calcium salts. Part II. The solubility of tertiary calcium phosphate in salt solutions and biological fluids; Part III. The solubility of calcium carbonate and calcium phosphate under various conditions,* Jour. Biol. Chem., vol. 71, no. 3, p. 783-846.

Shepard, F. P. (1941) *Nondepositional environments off the California coast,* Geol. Soc. Am., Bull. 52, p. 1869-1886.

———, and Emery, K. O. (1941) *Submarine topography off the California coast: Canyons and tectonic interpretations,* Geol. Soc. Am., Spec. Paper 31, 171 pages.

Thompson, T. G., and Robinson, R. H. (1932) *Chemistry of the sea,* Nat. Res. Council, Bull. 85, p. 112.

Trask, P. D., *et al.* (1932) *Origin and environment of source sediments of petroleum,* Am. Petrol. Inst., Gulf Publishing Co., Houston, Texas, p. 120.

Twenhofel, W. H. (1932) *Treatise on sedimentation,* 2d ed., Baltimore, Williams and Wilkins Co.

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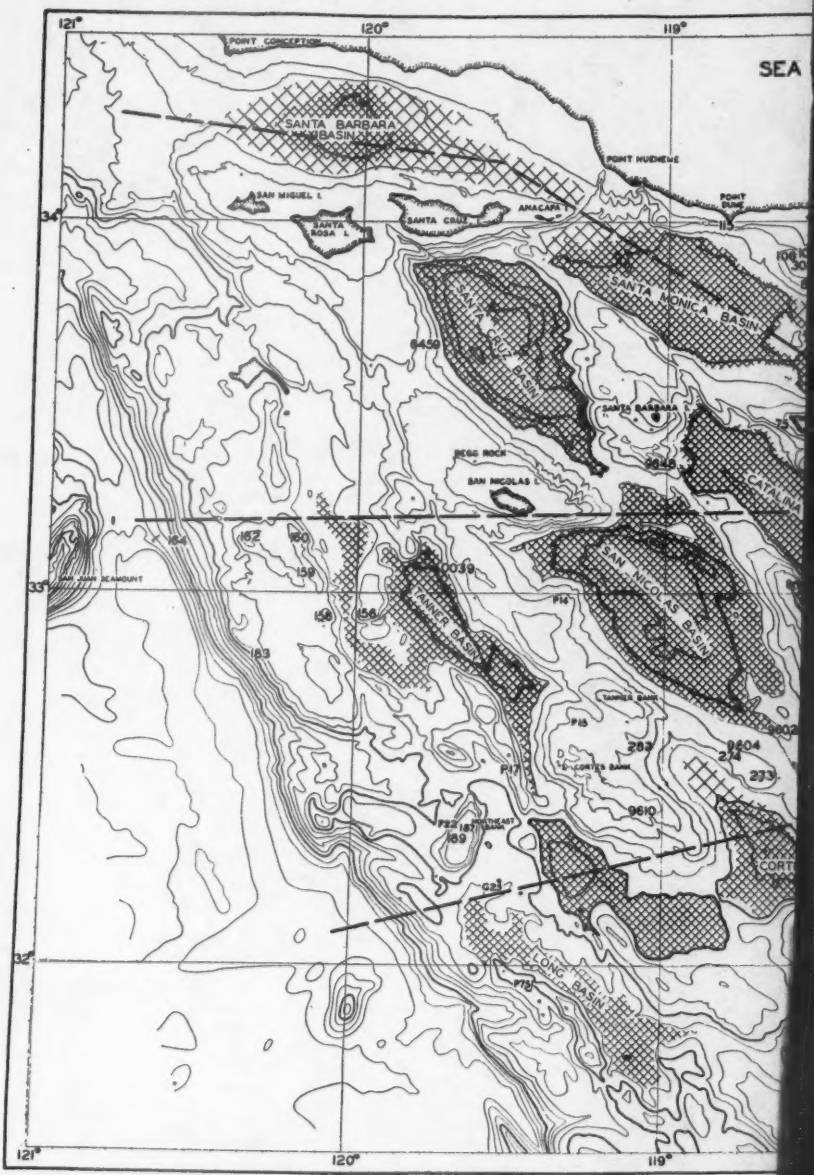
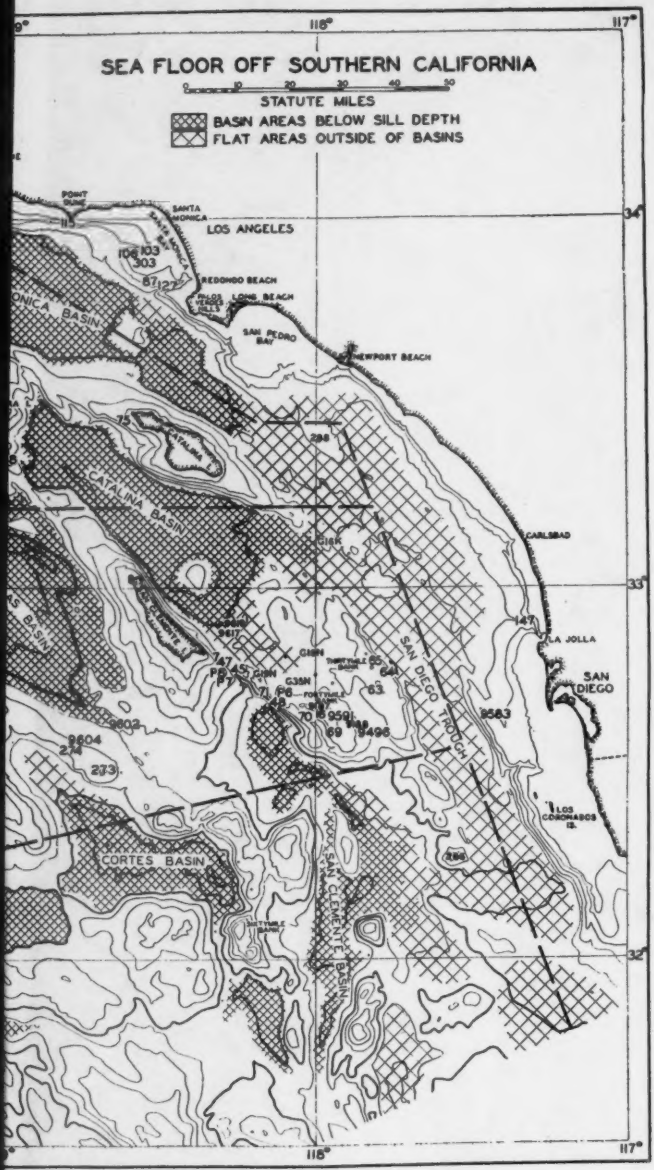


PHOSPHORITE LOCALITIES OFF SOUTHE

Showing the relation of phosphorite to the topographic highs. Contour interva numbers refer to samples containing large fragments of phosphorite and small num or phosphatic coverings on other rocks. Dashed lines indicate arbitrary divisions abundance. (*See* text).



FF SOUTHERN CALIFORNIA

Contour interval 900 feet with the 300 foot contour added. Large
e and small numbers to samples containing only phosphatic oölites
trary divisions of continental borderland related to phosphorite

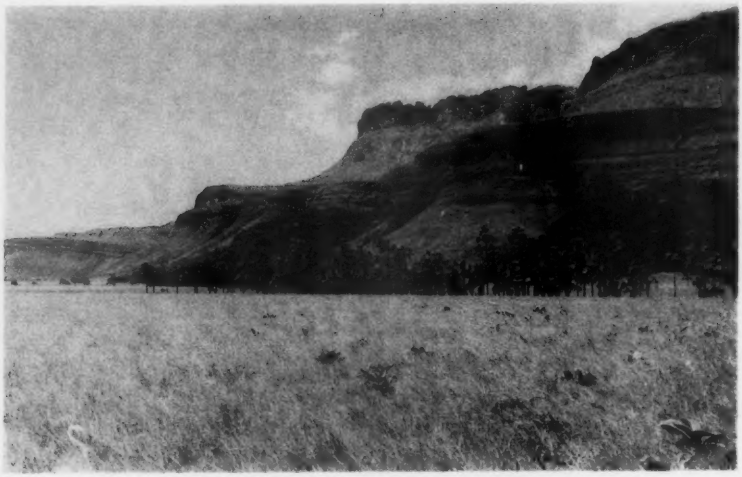


FIGURE 1. AUYÁN-TEPUI FROM THE SOUTH

FIGURE 2. MOUNT RORAIMA (SOUTH SIDE)
Upper part composed of sediments; lower slopes of quartz gabbro.

TYPICAL OUTCROPS OF THE RORAIMA FORMATION

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GEOLOGY OF SOUTHEASTERN VENEZUELA

BY V. M. LÓPEZ, E. MENCHER, AND J. H. BRINEMAN, JR.

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ABSTRACT

In southeastern Venezuela, in the region of the Gran Sabana, there outcrops the Roraima formation of conglomerates, sandstones, and shales, with interbedded thin tuffs. The greatest thickness, 2400 meters, at Mount Auyán-tepui, represents merely an erosion remnant. Predominant reddish tints, cross-bedding, arkosic materials, lateral and vertical variations, and the complete lack of fossils indicate a terrestrial origin. Thick dikes, sills, and laccoliths of gabbroitic rock intrude the Roraima, as well as a few acid dikes probably related to the same magma. The sediments are highly silicified. The age of the Roraima is undetermined.

The Roraima formation was deposited on the peneplaned surface of the Archean basement of the Guayana Shield. In the Gran Sabana area, these old rocks outcrop only along the Brazilian border where they consist of red and gray silicified porphyries.

The Roraima beds are essentially horizontal, except for gentle arching due to the gabbroitic intrusions. The present topography of mesas and cuestas is controlled by an original northward dip of the sediments off the old basement combined with the structures resulting from intrusion. The drainage is controlled by stream piracy due to a gentle southward tilting of the older northward-dipping beds.

Gold and diamonds in small quantities are associated with the basal conglomerate of the Roraima series and are concentrated in the Recent alluvial deposits eroded from the conglomerate.

INTRODUCTION

LOCATION OF AREA AND GENERAL REMARKS

In the eastern part of Venezuela south of the Orinoco River lies the Venezuelan Guayana. With the adjacent areas of the Guianas and northern Brazil, it constitutes the "Guayana Shield". Due to its inaccessibility, it is one of the least known sections of the country, and, except for the gold-mining centers in the north, has been known only through the work of a few explorers. Outside of the mining centers the region is inhabited solely by a few scattered Indian tribes.

In recent years the Venezuelan Government has realized the need of a better knowledge of this area with a view toward its agricultural and economic development, and General E. López Contreras, then President of Venezuela, in December 1938, authorized an expedition to explore parts of the Guayana country, under the general direction of the Servicio Técnico de Minería y Geología, Ministerio de Fomento. The expedition was divided into two groups. One, under the direction of G. Zuloaga and M. Tello, studied the northernmost extent of the Guayana where the high-grade iron ores previously described by Zuloaga (1930; 1933) and Burchard (1931) occur. The second group, headed by S. E. Aguerrevere, V. M. López, C. Delgado, and C. A. Freeman, limited its work to the southernmost portion of the Guayana, centering its activities in the open plain locally called the Gran Sabana (Pl. 2) and the surrounding highlands. Geographic, geologic, meteorologic, agricultural, and other studies were made, and the results were published as Volume 19 of the Revista de Fomento (Zuloaga and Tello, 1939; Aguerrevere, *et al.*, 1939). Since then, much new work has been

done in the laboratory on the specimens brought back by the second group, and this paper will present the new information as well as review for American readers the more important conclusions published in the Revista.

The area covered by this paper is therefore limited to the southeastern part of the Venezuelan Guayana. Geologically and geographically it extends into British Guiana and Brazil, which bound it on the east and southeast, and, in Venezuela, is bounded on the west and southwest by the Caroní River, and on the north approximately by the 6°30′ parallel. (*See* Plate 2.) Its area is about 35,000 square kilometers and is more or less included within 4°30′ and 6°30′ N. Lat. and 60°34′ and 62°50′ W. Long. Politically, it constitutes the central part of the District of Piar and the central and southeastern parts of the District of Roscio in the State of Bolívar. The actual route of the expedition is shown in Plate 2. In addition, the complete area was surveyed from the air.

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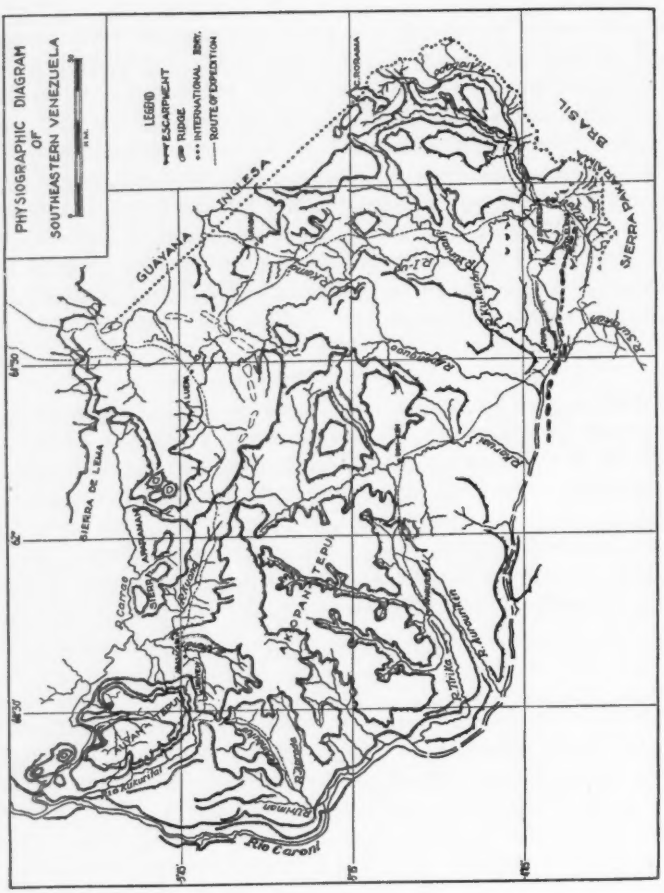
GEOMORPHOLOGY

TOPOGRAPHY

The southeastern Guayana area is divisible into two physiographic units—the gently sloping plain of the Gran Sabana (Pl. 2; Fig. 1) and the higher plateau regions surrounding it. The Gran Sabana area is characterized by a gently rolling topography, at an altitude of some 1100 meters in its northernmost limit south of Luepa, which slopes gradually southward toward the Caroní River and the Brazilian border where it is some 900 meters high. Most of it is underlain by the Roraima sediments and, as these dip north, a series of low cuestas facing south has been formed. The topography is somewhat complicated by the outcropping of a sill-like mass of gabbroitic composition about midway between Luepa and Uon-kén. As the igneous intrusions into the Roraima sediments have imposed a secondary structure of gentle warps and domes the mass between Luepa and Uon-kén


PHYSIOGRAPHIC DIAGRAM
OF
SOUTHEASTERN VENEZUELA
LEGEND
ESCARPMENT
RIDGE
INTERNATIONAL BDRY.
ROUTE OF EXPEDITION
GUAYANA
INGLESA
BRASIL
SIERRA PACARAIMA
SIERRA DE LEMA
Rio Caroni
Rio Orinoco
Rio Paragua
Rio Caura
Rio Aro
Rio Icabaru
Rio Yuruani
Rio Kukenan
Rio Karuai
Rio Yuruari
Rio Cuyuni
Rio Mazaruni
Rio Supamo
Rio Suapure
Rio Caronicito
Rio Paurai
Rio Yuruan
Rio Caroni

causes the sediments to the south to have a southerly dip and those just east of the valley of the Karuai an easterly dip. This, with the regional north dip of the beds farther to the south, has produced a gentle basinlike structure just east and northeast of Uon-kén. Subsequent erosion caused the cliffing of the edges of the basin so that, today, a large mesa faces the valley of the Aponguao to the east and southeast, of the Karuai to the southwest and west, and of a tributary of the Karuai to the north. On the whole, the Gran Sabana presents a contrast to the highland areas in that it has the appearance of a more mature topography with large rivers, such as the Karuai, Aponguao, and the Kukenán, meandering in broad valleys, some of which contain thick alluvium.

In the southernmost edge of the Gran Sabana, just north of the Brazilian border, acid porphyries which are part of an igneous complex underlie the Roraima sediments and extend into Brazil. Here the cuesta and plain character of the topography is lost, and a gabbro sill passing just south of Apoipó and extending beyond Santa Elena has produced a long, irregular, broken ridge due to differential erosion.

The highland regions may be divided into three subunits that owe their differences to structure. Separated from the Gran Sabana by the valley of the Karuai is the area of Western Plateaus or Mesas. These are broad and flat-topped, standing 2000 meters above the surrounding valleys. Most notable are Auyán-tepui and Akopán-tepui. (Tepui is a carib Indian word for hill.) The mesas are of the indurated Roraima sandstones, and structurally each mesa is saucer-shaped.

Separating the adjacent mesas of Auyán-tepui and Akopán-tepui is a gentle arch, now deeply eroded by the Ríos Ambutuir and Aprada in the southwest and by the Akanán in the north. These rivers have cut down through the sediments and into the gabbro intrusives, leaving only a small remnant of the sandstones known as Enguarapaima-tepui. The saucer-shaped mesas give the key to the origin of the topography of the Western Plateau zone. The sandstones, deep below the surface, originally dipped gently northward. Intrusion of gabbro later raised a series of low arches. Later, following uplift and stripping, the superimposed streams cut into the sandstones and exposed the gabbro in the arches. The present topography is the result.

Auyán-tepui is an irregular elliptical mesa, 35 kilometers long, 25 kilometers wide, and 2200 to 2400 meters high. On the south, facing the valley of the Uruyén, is a scarp in the form of three benches, each with a drop of about 600 meters. The lowest bench connects over the eroded arch with Akopán-tepui. The surface of Auyán-tepui is broken

into blocks by joint systems that may have had their origin in the intrusion of the gabbros and aided in the erosion of the scarps. During the rainy season much water accumulates on the mesa surface, and this is drained by several tributaries of the Río Carrao and one of the Kukurital. The waters of these tributaries tumble over the escarp-

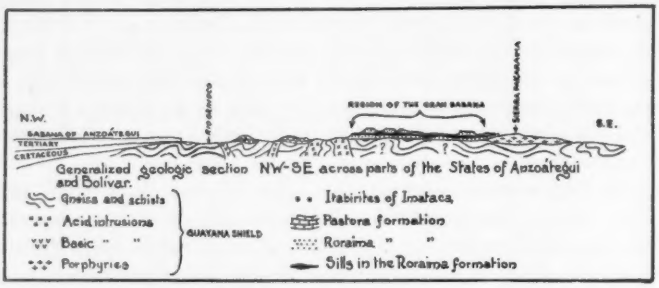


FIGURE 2.—*Generalized geologic section across parts of the States of Anzoátegui and Bolivar*

ment in imposing falls, and the recently discovered Angel Falls in the north-central part, where the three benches merge as a single scarp, probably ranks as the highest in the world, with a drop of at least 1200 meters.

Akopán-tepui is similar topographically and structurally to Auyán-tepui but is somewhat larger, covering an area of 2100 square kilometers. Except in the north, where there are two benches, the bounding scarp is simple and some 1000 meters high. The mesa has been almost completely bisected by the Río Tírika which, having its source at the top of the mesa, flows in a southerly direction to join the Río Aparurén at the foot of the escarpment and finally ends as a tributary to the Caroní. Other smaller streams also drain the surface of the mesa and flow in deep gorges.

In Auyán-tepui, Akopán-tepui, and, indeed, in all the smaller, similar units of the Western Plateau region, the dip of the underlying sandstones scarcely ever exceeds 5°.

Bounding the Gran Sabana on the east, the second highland subunit might be termed the region of the Eastern Mesetas which extends from the Río Kamoirán on the north to Mount Roraima and closely follows the Venezuela-British Guiana boundary. It is characterized by a series of high (2600-2800 meters) but relatively small mesas cut from a capping of resistant Roraima sandstone and quartzite beds overlying a laccolithiclike mass of gabbro. Generally, the surface of the sedi-

ments is penetrated by deep joints enlarged by erosion. The drop from the mesa surface to the gabbro is usually precipitous, there being but a single scarp (900 meters at Mount Roraima) (Pl. 1, fig. 2). Farther west is a scarp formed by the gabbro itself, but this is much less impressive than that originating in the sediments. Beneath the gabbro scarp lie the softer Roraima sandstones of the Gran Sabana.

The Eastern Meseta region forms the divide between three systems of drainage: (1) southwestward across the Gran Sabana to the Caroní; (2) southeastward into Brazil; (3) northeastward in British Guiana to the Mazaruni. Thus, some of the most powerful rivers of northern South America have their headwaters in the Eastern Mesetas. A number of these tributaries have their origins on the tops of the high mesas and drop onto the gabbro in falls that rival those of the Western Plateau region.

North of the Gran Sabana is a high saucer-shaped basin with its center near Luepa. In general, it does not lie much above the 1300-meter level of Luepa. This area constitutes the third highland subunit. The Luepa Basin is cut essentially upon the Roraima sediments (although gabbros do outcrop at a number of localities), but the more or less circular escarpment is underlain by the gabbros. In turn, the gabbros overlie a low, broad bench of sediments below which is a lower gabbro. The lower sandstone bench, as well as the underlying gabbro, is the northernmost part of the Gran Sabana proper. Structurally, however, the lower bench is more closely related to the Luepa Basin for it represents the edge of a shallow basin in which the Luepa Basin nests.

Though lower and broader than the Western Plateaus, the Luepa Basin is genetically very similar. The structure again is dependent upon the thick gabbro intrusives, here occurring at several horizons, with the sandstones and conglomerates dipping very gently from the periphery of the basin.

Between the Luepa Basin and Auyán-tepui are a series of small mesetas similar in general structure to the mesas of the Western Plateaus.

DRAINAGE

Even in their broadest features, the streams that drain the southeastern Venezuela region are completely unadjusted to the rock structure. The larger rivers, excepting the Caroní and the Carrao, flow southward and, in general, against the dip of the strata. Moreover, many streams, such as the Aponguao, have extremely abnormal courses

due, in large part, to piracy. Uplift is evidenced by the incising of the streams in the sandstones and by the apparently very recent alluvial terraces.

One of the most striking features of the rivers is that they flow, perhaps in a meandering course, for a considerable distance along a sandstone level and then suddenly drop, in falls of imposing height, onto the next lower resistant sandstone level. Where this occurs, the streams cut across the structure. Such falls occur in the Aponguao about 18 kilometers south of Luepa, in the Kukenán just above its confluence with the Arobopó, in the Tírika 5 kilometers below Aparurén, in the Yuruani near where it is crossed by the Luepa-Santa Elena trail, and even in the Caroní a short distance below the mouth of the Urimán. The main streams, therefore, show a more or less obsequent relationship to the sandstone structure.

A complete interpretation of the drainage must await a careful mapping and geomorphologic study. However, some general conclusions may be advanced from the data in hand. In the northern part of the area, in the headwaters and upper tributaries of such rivers as the Karuai and Aponguao, there is a notable lack of conformity between the stream course and its general southern direction of flow. For instance, the Aponguao rises in the northern part of the Luepa Basin where it flows almost due eastward practically to the British Guiana border. Here it reverses its direction and assumes a southwestward meandering course. South of Luepa it meets a large eastward-flowing tributary. The main stream, still meandering, again changes and flows for about 5 kilometers toward the southeast only to revert to the southwest and then to the south where it falls over an escarpment onto the gabbro. The course becomes more easterly and then southeasterly up to the confluence with the Kamá, and from then on a westerly and southwesterly flow prevails as far as the mouth of the river at the Caroní. To consider this the normal course of a stream seems hopeless; but to interpret the changes in flow as a result of piracy and reversal of original flow appears much more logical. It is not difficult to conceive of the Aponguao up to its assumption of the southeasterly course just south of Luepa as flowing northeast into British Guiana or think of the Kamá as flowing northeastward into one of the three tributaries of the Kamoirán near Uradai. Thus, the present river can be broken up into a number of older northeastward-flowing portions now joined by later southward-flowing ones.

It seems likely that the present headwaters of the Karuai (which even today flow northwestward) had a northward course, perhaps joining with the tributary of the Carrao that flows in a trench at the

eastern edge of the Sierra Aparamán. Other parts of the present river probably also once had a different course. The Carrao itself shows evidence of capture north of the eastern part of Auyán-tepui.

In general, therefore, the early drainage of the region was to the north and very likely consequent upon a fairly gentle northward-dipping structure. The stage of dissection in this early cycle can only be conjectured, but the well-developed meanders in the upper courses of some of the rivers would seem to indicate that at least maturity had been reached.

Later, uplift and tilting took place. While the latter was not enough to reverse the direction of dip, it was sufficient for the development of new streams as obsequents from the south which later captured parts of the old northerly system.

The related problems of the reasons for the uplift and tilting and the time of its occurrence cannot be solved at present.

Intrusion of the large gabbroitic masses may have caused the movements; if so, the present geomorphology depends wholly upon the gabbros. However, the evidence is just as much in favor of the occurrence of the intrusions before the uplift. At any rate, the exposure of the present structures (gentle arches and basins in the plateau areas and a more or less regional northward dip in the Gran Sabana) by uplift and erosion is responsible for the cuestas and high mesas.

This erosion into benches has also affected the courses of some of the streams. Many flow for considerable distances along one bench before falling onto the one below, as in the western branch of the Río Tírika that joins the main stream near Aparurén. Farther along, the Tírika itself follows a bench up to its confluence with the Caroní. The Caroní, the main river of the area, has its course along the benches, being incised between the cliff and the edge of the bench and occasionally falling to the next lower level in a series of cataracts or an imposing falls. Its change in position along the benches can be seen near its junction with the Kukurital where it has abandoned an old channel to reach a lower level more rapidly. Ultimately, the waters of the Caroní are poured into the Orinoco, but this is far beyond the northern limits of the area under consideration.

STRATIGRAPHY

The stratigraphy may be summarized as follows:

Alluvial deposits Quaternary	Gravel, sands, and clays. Diamonds.
Roraima formation Age?	Sandstones, silicified sandstones (quartzites), conglomerates, and shales, white to pale purple; thin beds of green and red jasper. Sills, dikes, and laccoliths of gabbro and diabase. A few acid dikes. 2400 meters minimum thickness. Basal conglomerate; pebbles of porphyries and itabirite. Diamonds? 10 meters.
Basement complex (Pre-Cambrian Shield)	Red and gray acid porphyries in southern Gran Sabana and Sierra Pakaraima. Gneisses, mica schists, granites, and other crystalline rocks, cut by basic intrusives, north of Gran Sabana.

DESCRIPTION OF THE ROCKS

BASEMENT PORPHYRIES

General statement.—While farther to the north the Archean basement rocks are typical, being crystalline metamorphics and granites cut by basic intrusions, in the area only fine-grained porphyries crop out. These occur at the southern edge, at Santa Elena, where they are red, and in the Sierra de Pakaraima, where they are gray. The two types interfinger throughout most of their extent and are quite similar except for an abundance of iron oxide in the former. Both have phenocrysts of plagioclase and more rarely of quartz. Both have undergone a later and almost complete silicification, and only remnants of the original texture can be seen.

Red porphyry.—This type is intensely red with small black spots irregularly distributed over the surface. Macroscopically the texture is fine, and few phenocrysts are visible.

Microscopically the texture is porphyritic. The relatively few phenocrysts make up 5 to 10% of the rock and vary from a fraction of a millimeter to 3 millimeters long, with an average of about one millimeter. The phenocrysts are acid andesine and quartz with the former predominating. All the phenocrysts show evidence of corrosion and, in addition, most of the plagioclase is sericitized. The groundmass is exceedingly fine-grained. The greater part consists of quartz, sericite, and limonite, the last of which is very abundant and gives color to the rock. In places, a now silicified original devitrification texture of matted minute crystal

laths can be discerned. The whole of the rock has been well silicified, and quartz veinlets abound in microscopic irregular cavities.

The red porphyry is evidently a silicified acid volcanic, but the high degree of silicification prevents exact classification.

Gray porphyry.—Macroscopically, the rock is dark gray, almost black, with distinctly visible white phenocrysts. Microscopically, the texture is porphyritic with an exceedingly fine groundmass. The phenocrysts constitute 10 to 30 per cent of the rock and vary in length from 1 millimeter to 4 millimeters, the average being 2 millimeters. The phenocrysts consist almost wholly of plagioclase (basic oligoclase or acid andesine) which has been partly altered to sericite and calcite, but a few quartz crystals are present. All the phenocrysts show corrosion. In certain portions of the groundmass, a now silicified microspherulitic structure can be seen which is probably original in the rock but which may have been introduced by silicification. Minerals of the groundmass are quartz, feldspar microlites, and secondary minerals such as magnetite, epidote, chlorite, and sericite.

The rock has been almost completely silicified, and innumerable quartz veinlets are present.

RORAIMA FORMATION

Resting directly but unconformably upon the Basement complex is a thick series of almost horizontal conglomerates, sandstones, and shales varying from white to red and gray, with occasional thin beds of red and green jaspers, all intruded by great masses of quartz gabbro.

The series begins with a basal conglomerate, which crops out near Santa Elena. The conglomerate varies from gray to pink. It is not massive but occurs interstratified with sands, in a total thickness of about 10 meters. The larger pebbles (a fraction of 1 centimeter to 12 centimeters long) are angular to well rounded and are composed of quartz, porphyry (from the basement porphyries), and of a sandstone made up of alternating laminae of quartz grains and hematite, showing soft-rock deformation, and suggestive of the itabirites of the Brazilian Serie de Minas and the Venezuelan Imataca formation. In the finer-grained matrix, quartz is the most abundant mineral. It is angular and, although quite a few of the grains show irregular inclusions (probably very minute cavities), the majority are quite clear. Quartz with inclusions similar to those mentioned above characterizes the veinlets in the Archean porphyries, and such a source, when taken in conjunction with the porphyry pebbles, seems highly probable. The clearer quartz may have originated in the phenocrysts of the porphyries. Magnetite, in small rounded grains averaging .015 mm^2 (associated quartz measures double), is very

abundant along certain layers. Other primary minerals are ilmenite (now changed to leucoxene), muscovite, and rounded zircon. Secondary minerals include hematite, limonite, and sericite. All specimens are highly silicified.

Above the basal conglomerate lies a minimum of 2400 meters of interbedded sandstones, conglomerates, and shales, converted in many places to quartzites and quartzite conglomerates. The sandstone is the most important, the shales and the conglomerates representing merely grain size variants. Thin beds of jasperized tuffs occur at many horizons. Macrostructures consist of frequent cross-bedding and occasional ripple marks. The bedding on the whole is good, but the size of the individual layers varies greatly. Lateral and vertical changes in grain size are common.

The lowest bench at Auyán-tepui shows:

4. Quartzite conglomerate with well-rounded pebbles of quartz and red, green, and black chert in a matrix of quartz and feldspar; 20 meters; overlying
3. Sandstone, purplish-gray, medium-grained; 90 meters.
2. Sandstones and shales, red, micaceous, with concretionary nodules; 60 meters.
1. Sandstones, soft, light-gray, pink, and pale-purple, micaceous, with thin layers of gray and pale-purple sandy shales; 50 meters.

At Mount Roraima, 500 meters of sandstones occur above the laccolith. At the contact, the sandstone has been converted into a light-gray quartzite. In a number of places, the quartzite contains bands of clear quartz which undoubtedly represent introduced material from the later stages of differentiation. Above the contact are cross-bedded, pink, medium-grained sandstones completely lacking in quartz veins or other macroscopic evidence of introduction of materials. Microscopic examination, however, shows a regrowth of the original quartz grains, and actually these higher sandstones should also be classified as quartzites. Conglomerates are absent.

Near Santa Elena the series consists of silicified conglomerates, quartzites, sandstones, and shales. Quartz is by far the most important mineral and has angular to subrounded boundaries, the better-rounded grains being more typical of the larger grains of the conglomerates. They vary from a small fraction of 1 millimeter in the shales to about 14 centimeters in the conglomerates. Some of the quartz is clear, but a large proportion contains the irregular inclusions previously discussed under the basal conglomerate and the porphyries. Very minute rutile hairs are also common, and many grains contain both rutile and the irregular inclusions. Occasionally apatite inclusions are also associated with the irregular inclusions. Chlorite inclusions are rare. Some of the quartz is of vein

origin. Later regrowth of the quartz grains or pebbles is common, but, in many instances, the original grain boundaries are still visible. In some specimens the quartz borders are very irregular due to replacement by sericite.

Much less important among the detrital minerals is acid plagioclase which occurs in small angular grains, but in a few specimens the plagioclase constitutes 5 per cent of the rock. This plagioclase may have its origin in the phenocrysts of the basement porphyries. Muscovite is usually present, but in very small proportion. Some is detrital, and some secondary.

One specimen shows good orientation of the muscovite flakes. Magnetite, in small rounded grains, occasionally becomes important in certain horizons, and one ferruginous layer in quartzite southeast of Santa Elena consists wholly of magnetite and yellow zircon. Zircon, too, occurs as one or two small rounded grains in most specimens except for local concentrations. A few grains of sphene are sometimes present also. Hematite makes up a considerable proportion of the red sediments and generally has two types of occurrence—one, as a detrital mineral in small grains; the other, as a secondary mineral surrounding grains or altering them, and along cracks in the rock. Detrital rock grains present are of vein quartz and porphyry. The smaller porphyry grains are angular. Pebbles or grains of crystalline igneous or metamorphic rocks such as are common in most places in the Archean basement are completely absent from the sediments.

Secondary minerals consist of hematite, limonite, sericite, and chlorite. Late alteration effects on the rocks may be summed up as silicification (quartz veinlets are present in some of the specimens) and sericitization. The silicifying solutions were probably those that arose in the later stages of the differentiation of the gabbroitic magma.

While the sediments in the area between Santa Elena and Luepa are quite similar to those just described from the Santa Elena region, enough difference exists to warrant a description. They consist of silicified conglomerates, quartzites, silicified arkoses, fine-grained red sandstones, and red shales. The quartz shows original rounded boundaries in most specimens except for the finer varieties. Again irregular and rutile inclusions are abundant, while apatite inclusions are rare. The most notable difference between these sediments and those described from Santa Elena lies in their higher percentage of feldspars. Rounded grains of plagioclase, orthoclase, and microcline are present in many specimens and in a number of cases make up 10 to 15 per cent of the total minerals present. Since a considerable amount of the feldspar has evidently been altered and removed, it is probable that 20 per cent is closer to the original content.

Thus, many of the sandstones must be considered arkoses. It would be interesting to try to analyze the reason for the highly feldspathic character of these rocks as compared with those of Santa Elena, but such a study would be of value only if both series of sediments came from the same horizon, and the nature of the field work precluded such careful sampling.

Zircon and sphene again are present as minor constituents, and hematite shows its usual character in the red sediments. Muscovite is rare. Detrital rock grains and pebbles consist of a very fine-grained silicified rock (probably porphyry), vein quartz, and, in one of the specimens, of a veined quartz grain. Secondary minerals are hematite and sericite, and practically all the specimens show silicification and sericitization.

Specimens from near Luepa do not differ materially from those already described, but one specimen contains pyrite in irregular grains and a calcite cement along certain layers. The calcite is well-aligned in that all of it extinguishes at the same position. This, however, appears to be a depositional feature.

The sediments from the first bench at Auyán-tepui are also similar to those from the other parts of the region. Formerly unnoted features shown by these rocks are the presence of introduced plagioclase with regrowth of some of the original plagioclase boundaries; an alignment, probably depositional, of muscovite in some of the shale horizons; and indications of soft-rock deformation in a few specimens.

To date, no fossils have been encountered. Moreover, the arkosic nature of much of the material, the presence of reddish and pinkish sediments, the frequent lateral and vertical changes of sandstone or shale through conglomerate, the frequency of cross-bedding, and the occasional ripple-marking all point strongly toward a terrestrial origin. The well-rounded quartz grains in many of the specimens indicate a distant source. For an environment of sedimentation, the present "llanos" deposits of the Orinoco River are suggestive, *i. e.*, an intermingling of stream deposits in a large basin at a considerable distance from their source and under somewhat dry conditions. Undoubtedly, some of the material came from the south, for the formation contains porphyry pebbles from the more southerly outcropping basement porphyries. The porphyry pebbles are most abundant in the lower part of the series, especially in the basal conglomerate, and are probably derived from a more immediate source than the later beds. As already mentioned, the basal conglomerate also contains pebbles of a rock resembling itabirite. However, itabirites outcrop both to the north in the Imataca formation on the northern limit of the Venezuelan Guayana

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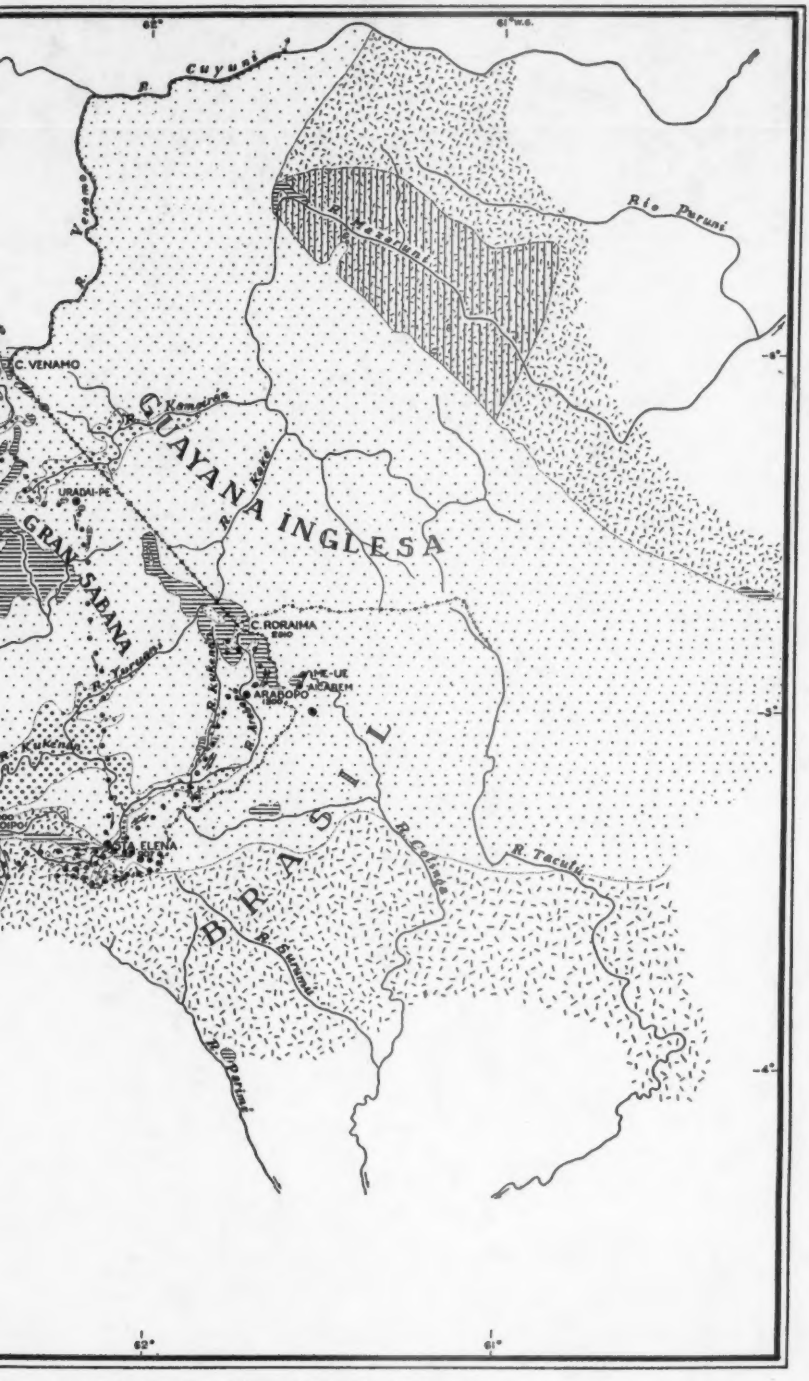

THE GRAN SABANA AND ADJACENT REGIONS

(along the southern bank of the Orinoco River from Ciudad Bolívar to the Atlantic Ocean) and to the south in the Proterozoic Serie de Minas of northern Brazil. If the southern derivation of the porphyry pebbles be taken as an indication of the source of the itabirite pebbles, it would still not necessarily localize the origin of the latter in one of the present-day occurrences in Brazil, since there is no evidence to disprove the former continuity of the Brazilian deposits with those in Venezuela or, indeed, to deny a much greater original areal extent of the itabirites. However, throughout long geologic periods previous to the deposition of the Roraima sediments the Guayana area was undoubtedly exposed to erosion, and it may well be that the present itabirite outcrops approximate their extent during the Roraima deposition.

Another interesting feature that may be of importance in determining the source of the deposits is the apparent association of diamonds and gold with the basal conglomerate of the Roraima formation. Although neither diamonds nor gold has actually been found in the basal conglomerate itself, both in British Guiana and in Venezuela (in the mining region of the valley of the Surukún, near Apoipó) the deposits of value are found in the recent alluvial sediments that have their origin in the erosion of the basal conglomerate. Thus, it would appear that the basal conglomerate itself was an ancient placer deposit. Such a diamantiferous conglomerate suggests the Sopa formation of the Proterozoic Serie de Lavras of the State of Minas Geraes of Brazil (Moraes, 1934, p. 21-34; 1937, p. 13-14), and the Brazilians correlate the Roraima beds with their Serie de Lavras. At any rate, the only existent source for the diamonds is in the Proterozoic formations of Brazil. Gold occurs in the Brazilian Serie de Minas and, again, therefore, a southern source is indicated.

BASIC INTRUSIVES

General statement.—Intruding the Roraima sediments (Pl. 2) in the form of dikes, sills, and laccoliths are thick masses of a dark-gray igneous rock varying from quartz gabbros and quartz diabases, through gabbros and diabases, to olivine diabase. (For chemical analyses, *see* Table 1.) All are compositional variants of the same magma, but the quartz gabbro is by far the predominating type. In many places they show a chilled border on contact with the sediments.

Quartz gabbro.—The quartz gabbro, which occurs in all parts of the region, is extremely uniform in texture and composition. Macroscopically, it is medium-grained, compact, dark-gray, with small white spots; microscopically, it is hypidiomorphic-granular, varying, in some cases,

to intersertal. There is frequently a minor micropegmatitic or myrmekitic structure that is interstitial with relation to the rest of the rock.

The essential minerals are labradorite, approximately Ab_{45} An_{55}, and

TABLE 1.—*Chemical analyses of the basic igneous rocks of the Gran Sabana*

	1.	2.	3.	4.	5.	6.	7.	8.
SiO_2	54.06	52.32	52.28	53.87	52.30	51.98	52.97	51.88
Al_2O_3	15.12	15.46	14.72	16.89	11.18	14.89	8.48	16.57
FeO	10.34	10.78	10.17	9.55	10.08	9.11	10.97	7.78
Fe_2O_3	1.18	1.42	0.62	0.81	0.81	0.34	0.74	1.06
TiO_2	0.79	1.44	0.73	1.47	0.72	0.73	0.72	0.58
MnO	0.26	0.26	0.09	0.16	0.19	0.14	0.14	0.12
CaO	8.99	9.66	10.72	9.48	8.52	11.12	8.30	11.83
MgO	5.69	5.12	8.13	3.73	14.26	9.54	15.88	7.22
K_2O	0.82	0.63	0.39	1.11	0.21	0.37	0.12	0.48
Na_2O	1.80	2.02	1.44	2.19	1.03	1.59	1.06	1.86
H_2O-	0.24	0.08	.....	0.05	0.17	.....	0.12	0.08
H_2O+	0.45	0.58	0.28	0.38	0.36	0.36	0.14	0.32
CO_2	.....	0.14	0.12	Tr.	0.28	Tr.	0.35	Tr.
P_2O_5	0.06	0.13	0.11	0.12	Tr.	Tr.	Tr.	0.09
	99.80	100.04	99.80	99.81	100.11	100.17	99.99	99.87
Specific gravity	2.97	2.93	2.71	3.02	3.06	3.06	3.04	3.06

1. Quartz gabbro from sill near Kamarata
2. Quartz gabbro from sill on road to Luepa from Kamarata
3. Fine-grained quartz gabbro from laccolith near Maparutá
4. Quartz gabbro from Kanaimé sill south of Santa Elena and Uairén
5. Quartz gabbro from Kanaimé sill south of Santa Elena and Uairén
6. Gabbro from laccolith near Luepa
7. Basic gabbro from Kanaimé sill south of Santa Elena and Uairén
8. Olivine diabase from Roraima

TABLE 2.—*Mineral analyses of the basic igneous rocks of the Gran Sabana*

	1.	2.	3.	4.	5.	6.	7.	8.
Labradorite	63	66	46	76	48	52	34	56
Augite	24	22	47	15	49	43	64	40
Uralite, Biotite, Chlorite	10	6	1	1	1	3	1	2
Magnetite	1	4	5	2	1	2	1	2
Myrmekite, Quartz	2	2	1	6	1	0	0	0
Total	100	100	100	100	100	100	100	100

augite. The former occurs as subhedral crystals, at times almost euhedral, averaging 1 millimeter in length. It is also found as blebs and crystals included in the augite. The plagioclase is altered, in part, to sericite, epidote, and kaolin.

The augite, usually diallage, is present in anhedral to subhedral grains from a fraction of a millimeter to 1.5 millimeters. In the diallage, the (001) parting is well developed. Uralitization of the augite boundaries is common and due to hydrothermal alteration. Alteration of the augite to chlorite is also frequent.

Accessory minerals are quartz, orthoclase, biotite, and apatite, and graphic micropegmatitic and myrmekitic intergrowths of quartz and orthoclase.

The biotite is the common brown variety showing strong pleochroism. It is largely chloritized with the production of secondary magnetite. It occurs in flakes replacing augite or uralite and was probably the first mineral introduced by the myrmekitic and micropegmatitic solutions.

Apatite is present in long, slender, euhedral prisms with characteristic cross-fractures.

The occurrence of the intergrowth structures is the most interesting feature of the quartz gabbro. Both types, myrmekite and micropegmatite, occur in the same sections. They are always interstitial to the plagioclase and augite and hence are later than either. In some cases the quartz of the intergrowth is in optical continuity with free quartz, which also occurs interstitially; in others, the free quartz of the rock takes the external form of the intergrowth material. Thus it appears that the myrmekite and micropegmatite were simultaneous in crystallization with a part of the free quartz but that the quartz deposition continued after the termination of the intergrowth crystallization, *i.e.*, after all the orthoclase components of the solution had been used up.

Orthoclase occurs almost invariably as a part of the intergrowth structure, but there are a few grains of free orthoclase.

Secondary minerals are magnetite (alteration product of biotite and augite), sericite (as alteration of the plagioclase), chlorite (as alteration of augite and biotite), uralite (as hydrothermal alteration of augite), and calcite (as alteration of plagioclase).

The order of crystallization of the more important minerals is labradorite, augite, uralite, biotite, myrmekite and micropegmatite, orthoclase and quartz, quartz.

Mineral compositions of the quartz gabbro are given in Table 2, columns 1 to 5.

Quartz diabase.—The quartz diabase represents merely a textural change from the hypidiomorphic of the quartz gabbro to diabasic. Many of the quartz gabbros studied microscopically showed variation toward ophitic textures, but only occasionally is this texture sufficiently developed to warrant calling the rock a quartz diabase.

Gabbro.—True gabbros are rare. One such occurrence is at El Tigre from the sills surrounding the Luepa Basin. The texture is hypidiomorphic-granular, and the essential minerals are labradorite and augite. The only accessory mineral is a small quantity of apatite. This rock is similar to the quartz gabbro except for the presence of quartz and the acid intergrowths in the latter. It would, therefore, appear to be a product of the same magma but to have been unaffected by its later acidic differentiates (Tables 1, 2, cols. 6, 7).

Diabase.—On the southeast flank of Mount Roraima, in the valley of one of the tributaries to the Río Kukenán, there is a diabase sill. The rock is fairly coarse-grained, compact, and dark-gray; under the microscope it is diabasic. Fresh euhedral labradorite makes up 50% of the rock. Large subhedral to anhedral crystals, some as much as 4 millimeters long, of augite enclose the plagioclase. A polysynthetic twinning is usually present. In some cases the augite has been replaced by another unidentified pyroxene giving the augite a micrographic appearance. The accessory mineral is biotite partly altered to chlorite and magnetite.

At an altitude of 1920 meters on Mount Roraima is a diabase sill containing about 5% olivine (Tables 1, 2, col. 8).

These diabases, too, are undoubtedly derivatives of the quartz-gabbroitic magma.

ACID DIKES

Acid dikes cut both the basic intrusives and the sediments in a few places. They are fairly uniform in composition and texture and might be classified as granophyres. The dikes are coarse-grained and light-colored and in thin section show a hypidiomorphic texture strongly modified by micropegmatitic structure. At times, the micropegmatite becomes the dominating texture.

Essential minerals are micropegmatite, perthite, plagioclase, and quartz. The micropegmatite consists of quartz and apparently a perthitic intergrowth of orthoclase and plagioclase. The perthite occurs almost wholly within the micropegmatite, but a few free grains are present. The plagioclase has an approximate composition of acid andesine, Ab_{70} An_{30}. It is not entirely fresh but shows alteration. Most of the quartz is in the micropegmatite, but a small percentage of free quartz does exist.

Accessory minerals are ordinary brown biotite, apatite in long slender prisms, and green hornblende.

Secondary minerals are magnetite (alteration of biotite and hornblende), chlorite (alteration of ferromagnesians), sericite (from the feldspars), and limonite.

In general, these acid dikes may be classified as microgranites or as hornblende-biotite granophyre. There are variations to quartz diorite, but these are special cases.

While it is impossible to state with certainty that the acid dikes are late derivatives of the quartz gabbroitic magma, the abundance of micropegmatite in both the interstitial material of the quartz gabbro and the acid dikes, as well as the similarity of the more acid minerals of both rock types, suggests that the acid rocks may be a continuation of the differentiation process that gave rise to the interstitial silicic materials of the quartz gabbros.

CONTACT EFFECTS

While contact effects are visible in the sediments, they must be ascribed to siliceous rather than basic solutions. For instance, the contact quartzite at Mount Roraima indicates the introduction of silica. At the same time, it indicates that the quartz and myrmekite represent a phase of the gabbroitic magma, for the silicified zone is closely related to the gabbro surface. The sandstones seem to have been a better host rock for the acid solutions than for the more basic ones; also, the greater fluidity of the siliceous solutions may have aided in the penetration of the sediments. Only in a xenolith of the sandstone in the gabbro are there contact effects, and these are limited to the introduction of biotite into the sandstone.

ALLUVIAL DEPOSITS

Along the courses of many of the present rivers are thick deposits of conglomerates, sands, and clays whose origin is in the sediments of the Roraima formation. They are predominantly white to yellow and red, but in a few localities there are fine-grained dark deposits that owe their character to vegetable material. The occurrence of diamonds and gold in the alluvial deposits south of Apoipó and their apparent relationship to the basal conglomerate of the Roraima formation have already been mentioned.

HISTORICAL GEOLOGY

AGE OF RORAIMA FORMATION

As the Roraima beds are unfossiliferous, only indirect methods can be applied to the determination of their age. Thus the Roraima rocks have been variously placed from the pre-Cambrian to the Eocene. While the reports on British Guiana (Brown, 1875, p. 14; Harrison, 1908, p. 23; Conolly, 1925, p. 23-24) discuss the problem of age they present no evidence for an age assumption. Branner (1919) colors the Roraima beds as Cretaceous on his map of Brazil but remarks (p. 230):

"Of the area about the head-waters of the Río Branco but little is known beyond what is to be had from the reports of Brown and Sawkins on the adjoining parts of British Guiana. The areas colored as questionable Cretaceous are merely areas of sandstone resting on the Archean. The rocks have as yet furnished no fossils."

More recent Brazilian reports (Oliveira, 1938) have, however, favored a Proterozoic age probably because of a lithologic similarity to the Serie de Lavras of northern Brazil and to the occurrence of detrital diamonds in the basal conglomerate of both series. While the presence of diamonds in both series may seem a strong point in favor of their correlation, one must remember that any conglomerate originating in the diamond-bearing Proterozoic rocks of Brazil or a derivative would be likely to carry diamonds. One need look no farther than the placer diamond deposits in the Recent alluvium in Brazil (erosion of original diamond-bearing intrusives in Serie de Itacolomy or secondary deposits in basal conglomerate of Serie de Lavras), British Guiana, and Venezuela (erosion of basal conglomerate of Roraima formation). Liddle (1928, p. 124) attempts a correlation with the Lower Cretaceous formations of the Venezuelan Andes, Coast Range, and Sierra de Perijá, and particularly with the Barranquín formation of northeastern Venezuela. Recently the tendency among geologists in Venezuela (Engleman, 1935, p. 782; Kugler, 1936, p. 1441; Hedberg, 1940) has been to correlate the Roraima formation with the 3000-meter thick Upper Triassic?-Lower Jurassic? (possibly Paleozoic) La Quinta-Girón beds of Venezuela and Colombia. From a lithologic viewpoint, this latter correlation seems best, for the La Quinta-Girón formation consists essentially of red to white sandstones, shales, and conglomerates with occasional tuffaceous beds. There are also basic intrusives and flows. Aside from fish remains there are no fossils. Thus, there is a great similarity between the Roraima beds and the La Quinta-Girón formation. Moreover, Gerth (1935, p. 202) shows a resemblance of the Roraima deposits to the Triassic of Brazil. The facts are insufficient, however, to define the age of the formation, and, indeed, tuffs and other evidences of volcanic activity are common in South America and the Caribbean region from the Triassic to the present.

SUMMARY OF GEOLOGIC HISTORY

The history of the southeastern Venezuela area begins in the pre-Cambrian with the Guayana Shield rocks. Very little is known about these rocks, except that they are highly crystalline gneisses, mica schists, porphyries, and granites. Following their formation and metamorphism in Archean times, the area was exposed to erosion and peneplanation. Farther south, in Brazil, three series of Proterozoic rocks as well as many Paleozoic and Mesozoic formations were laid down in

various places upon this peneplane, but in southeastern Venezuela there is no trace of any deposition between the time of formation of the peneplane and the deposition of the Roraima beds. This may mean that from pre-Cambrian time onward this part of the Brazilian Shield was a positive element above sea level or that overlying formations were completely removed before the Roraima cycle began. Whatever the early history of the region the Roraima sediments lie directly upon the Archean peneplane surface. The early Roraima deposits do not indicate lowering of this old surface but rather the rise of a near-by area, probably to the south, with the rejuvenation of the streams and even possibly with a change in the direction of flow so that a series of fast-moving streams flowed over the old surface and began to lay down coarse gravels in which the pebbles were derived from the newly risen area. These pebbles show their source to have been porphyries, such as now outcrop to the south, and itabirites, which today outcrop both south and north of the region. Also deposited in these conglomerates were occasional grains of diamond and gold, both of which probably had their source in the Proterozoic rocks farther south in Brazil. (The rich Venezuelan gold deposits of El Callao lie to the north, but the association of the diamonds with the gold points rather strongly toward a southern source for the sediments.) These early gravels probably did not travel far before their deposition, but, as the cycle continued, the area covered by the sediments increased, and the materials traveled a much greater distance to reach southeastern Venezuela. Meager evidence still points toward a southern source, for the majority of the quartz grains abound in microscopic, irregular cavities, and such a type of inclusion is also present in the quartz veining the gray porphyries. These later deposits are finer-grained on the whole than the basal sediments but show all typical structures of stream deposition such as cross-bedding, lateral and vertical variation of grain size, occasional conglomerates, ripple-marking, pink color, and abundance of feldspar. Many of the sandstones might be classified as arkoses.

During the deposition of the sediments the air would occasionally be filled with the finer materials erupted from volcanoes at a considerable distance from the basin, and a thin layer of tuff would result.

However, it is necessary to assume a slow sinking of the area to accommodate the more than 2400 meters of deposits laid down between the beginning and the end of the Roraima sedimentation. The Roraima deposits, therefore, represent a tremendous basin of interfingering alluvial sediments laid down in a slowly sinking area. The present "llanos" deposits of the Orinoco Basin perhaps demonstrate the conditions during

Roraima times, except that the present llanos climate is probably more arid.

The original extent and thickness of the deposits are speculative. Did they connect with the 3000-meter thick La Quinta-Girón beds to the west in the Andes? Did they continue to the north and interfinger with marine sediments of the Coast Range geosyncline in Venezuela and Trinidad? Were they joined to deposits farther south in Brazil?

At some time after the deposition of at least the remaining part of the Roraima sediments a magma of gabbroitic composition started to invade the deposits. If we assume a great original thickness for the Roraima sediments it is possible to imagine that the intrusion took place during the later stages of Roraima deposition, for the intrusions may have been sufficiently below the surface for the good crystallization of the magma. On the other hand, the probability is just as strong that the intrusions came in long after the end of the sedimentary cycle, and there may even have been later beds above the Roraima sediments that have now been completely eroded. This gabbroitic material forced itself along the layers in thick masses, causing a very gentle arching of the overlying sediments. A later phase of the same magmatic cycle brought in solutions rich in silica and resulted in the filling of interstices in the gabbro with quartz and myrmekite and the veining and silicification of the sediments. In the sandstones and conglomerates the original quartz grains were regrown, and some of the rocks were converted to quartzites. In most cases the feldspar was removed, but a few specimens show regrowth and introduction of plagioclase. The fine-grained tuffs seem to have made excellent hosts for the siliceous solutions and were changed to jaspers.

Possibly a still later cycle of acid intrusion took place as evidenced by micropegmatite dikes cutting the gabbros and diabases, but even these may be but a phase of a single intrusive period.

The arching of the sediments by the intrusive mass greatly affected the later topography and drainage. Before the uplift and exposure of the arched beds, the rocks at the surface had a gentle northerly regional dip off the old porphyries, and the streams, therefore, assumed a consequent northerly flow of dendritic pattern. Later uplift and tilting to the south took place with the capture of the old northward-flowing streams by younger southward-flowing obsequents, and the arched beds and intrusives were brought to the surface to be eroded into the present topography.

ECONOMIC GEOLOGY

Two minerals of economic value are found in the region: gold and diamond. While gold is common in many places in the Basement com-

plex in Venezuela and Brazil and also in the Lower Proterozoic Serie de Minas of Brazil, and while diamond occurs in metamorphosed intrusives in the Middle Proterozoic Serie de Itacolomy and in the conglomerates of the Sopa formation of the Upper Proterozoic Serie de Lavras of Brazil and in Recent sediments derived by erosion from these rocks, the almost total absence of such rocks in the area under discussion would seem to minimize the possibilities of encountering deposits of value. Nevertheless, gold has been found in small quantities at a number of localities, and diamonds and gold in commercial quantities are found together south of Apoipó in the valley of the Río Surukún. (*See* Plate 2.) All these occurrences are in placers resulting from the erosion of, presumably, the Roraima formation.

At present the only deposits being exploited are those in the Surukún Valley. The workings are limited to the upper parts of short ravines that are tributaries to the Surukún on its western side. These ravines originate on the eastern slopes of two sandstone mesas known as Paraitepui and Egua-tepui. The lower parts of the mesa slopes are composed of the basal conglomerate of the Roraima formation. At the foot of the conglomerate, the Basement complex porphyries outcrop.

The workings themselves are in the alluvial deposits filling the ravines close to the contact beween the basal conglomerate and the porphyries. The commercial deposits are undoubtedly associated with the conglomerate and come from its erosion. With the diamonds and gold are rounded pebbles and boulders mainly of the conglomerate and sandstone with a minor quantity of red porphyry and quartz. This relationship of the commercial deposits with the Roraima basal conglomerate has also been noted in British Guiana (Conolly, 1925, p. 8-9, 59, 88-92).

Exploitation of the area began in 1925, and it is estimated that up to the present over $500,000 worth of gold and diamonds has been mined.

Very small quantities of gold have been panned from the region known as the Kamarata Valley near Uruyén. The Venezuelan Government Expedition itself panned over 800 kilos of material from this area with a yield of only 4.43 grams of gold. The interesting feature of these deposits, valueless though they may be, is that they seem to be related to the intrusives of the Roraima formation rather than to the sediments. In this area, besides the quartz gabbros, there are acid dikes as well as quartz veins, and it is possible that the gold had its origin in the acid materials.

REFERENCES CITED

Aguerrevere, S. E., López, V. M., Delgado, C., and Freeman, C. A. (1939) *Exploración de la Gran Sabana,* Rev. Fomento, Venezuela, Año 3, N° 19, p. 501-542, 632-729.

Branner, J. C. (1919) *Outlines of the geology of Brazil to accompany the geologic map of Brazil,* Geol. Soc. Am., Bull., vol. 30, p. 189-337.

Brown, C. B. (1875) *Reports on the geology of British Guiana,* London.

Burchard, E. F. (1931) *The Pao deposits of iron ore in the State of Bolivar, Venezuela,* Am. Inst. Min. Metall. Eng., Tr., vol. 96, p. 355.

Conolly, H. J. C. (1925) *Report of the preliminary survey of the Mazaruni and Peruni diamond fields, British Guiana,* Econ. Geol. Survey, British Guiana.

Dalton, L. V. (1912) *Venezuela,* London.

Engleman, R. (1935) *Geology of Venezuelan Andes,* Am. Assoc. Petrol. Geol., Bull., vol. 19, p. 769-792.

Gerth, H. (1935) *Geologie Sudamerikas,* Pt. 2, Berlin.

Harrison, J. B. (1908) *Geology of the goldfields of British Guiana,* London.

Hedberg, H. D. (1940) *Mesozoic stratigraphy of northern South America,* unpublished paper presented at 8th Am. Sci. Congr., Washington, D. C.

Kugler, H. G. (1936) *Summary digest of the geology of Trinidad,* Am. Assoc. Petrol. Geol., Bull., vol. 20, pt. 2, p. 1439-1453.

Liddle, R. A. (1928) *The geology of Venezuela and Trinidad,* Fort Worth.

Moraes, L. J. de (1934) *Depositos diamantiferos no norte do Estado de Minas Geraes,* Dpto. Nac. da Prod. Min., Brazil, Bol. 3.

——— (1937) *Geología económica do norte de Minas Geraes,* Dpto. Nac. da Prod. Min., Brazil, Bol. 19.

Oliveira, A. I. de (1938) *Mappa geologico do Brasil e de parte dos paizes visinhos,* Min. da Agric., Serv. de Fomento da Prod. Min.

Zuloaga, G. (1930) *The iron deposits of the Sierra de Imataca, Venezuela,* Econ. Geol., vol. 25, p. 99-101.

——— (1933) *The geology of iron deposits of the Sierra de Imataca, Venezuela,* Am. Inst. Min. Metall. Eng., Tech. Publ. 516.

——— and Tello, M. (1939) *Exploración preliminar de la Sierra de Imataca,* Rev. Fomento, Venezuela, Año 3, N° 19, p. 398-430.

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NEW DEVONIAN (HAMILTON) CORRELATIONS

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ABSTRACT

Because of many easily recognized and restricted species, the fauna of the thin Centerfield formation of New York has considerable stratigraphic importance. The fauna is readily traced across New York from Chenango Valley to Erie County and appears in Ontario and Michigan. In Ohio, Indiana, and Kentucky, the fauna occurs with its important elements. In less typical aspect Centerfield fossils appear in northwestern Arkansas and possibly in West Tennessee. The Centerfield equivalent forms the top of the Hamilton in the Midwest, thus indicating a Marcellus-Skaneateles age for the underlying Hamilton beds.

INTRODUCTION AND ACKNOWLEDGMENTS

Correlations presented here for the first time enable detailed matching of the parts of the Hamilton group occurring outside of New York.

FIGURE 1.—*Map showing location of areas of Centerfield formation and its correlates*

(1) Emmet and Charlevoix counties, Michigan.
(2) Alpena County, Michigan.
(3) Thedford and Arkona, Ontario.
(4) Western New York.
(5) Wabash River area, Indiana.
(6) Lingle Creek, Union County, Illinois.
(7) Clark County, Indiana; Jefferson County, Kentucky.
(8) Plum Brook and Prout, Ohio.
(9) Chenango Valley, New York.
(10) Stroudsburg, Pennsylvania.
(11) Boyle County, Kentucky.
(12) Harpeth River, Tennessee.
(13) Eureka Springs-Harrison quadrangles, Arkansas.

Although generalities of the Hamilton of the Midwest are well known, no one has stated what parts of the group are there represented in terms of the New York section. Study of the Centerfield fauna now shows that no Moscow strata occur west of New York and the Appalachians and that the Ludlowville above the Centerfield is absent from most of the Midwest.

The writers studied the New York section individually but in 1934 jointly measured the sequences in southwestern Ontario and Michigan. In 1935 the joint studies were extended to Indiana and Illinois. In 1936 Cooper studied sections in southeastern and central Missouri. In 1937 Cooper studied parts of the Hamilton sequence in Pennsylvania with Dr. Bradford Willard. Cooper studied the eastern New York sequence and parts of eastern Pennsylvania in 1938. The authors acknowledge their indebtedness to the following local geologists: Mr. Charles Southworth of Thedford, Ontario; Mr. Guy Campbell of Corydon, Indiana; and Dr. Bradford Willard of Lehigh University.

HAMILTON KEY BEDS

The Hamilton group of New York forms a great westward-thinning wedge of clastics. Four thin calcareous layers, best developed in western and central New York, contain recurrent Onondaga types. These are the Mottville-Stafford, Centerfield, Tichenor, and Portland Point limestones. Although their faunas are similar they are easily separable in New York. The Centerfield is unique in containing certain faunal elements absent or excessively rare in the other limestones and identified at many places outside of New York.

CENTERFIELD FAUNA

Fossils most useful for correlation have a restricted vertical range but wide geographical distribution. The added quality of ease of identification makes these fossils ideal indexes. A number of such persistent species occurs in the Centerfield. The writers believe that the following are the most important: *Aveolites goldfussi, Favosites turbinatus, F. placentus, Prismatophyllum,* sp., *Megistocrinus* sp., *Dolatocrinus* sp., *Eleutherocrinus cassedayi, Camerophoria kernahani, Cyclorhina nobilis, Eunella attenuata, Camarospira* n. sp., *Pentagonia bisulcata, Parazyga hirsuta, Trematospira gibbosa, Spirifer venustus, S. sculptilis, Centronella impressa, Vitulina pustulosa,* and *Richina truncata.* Nowhere else in the Middle Devonian is this combination of genera duplicated. It is approximated in the *Meristella*-coral zone of the Marcellus, the Tichenor, and Portland Point, but, in each of these, important elements are missing or the proportions of the species are too unlike to cause confusion with the Centerfield.

Two important factors bearing on Centerfield species are: (1) Most of the species occur in other parts of the Hamilton, and (2) geographic expressions of the species occur. In the first instance *Centronella impressa, Vitulina pustulosa,* and *Spirifer venustus* all range as high as the

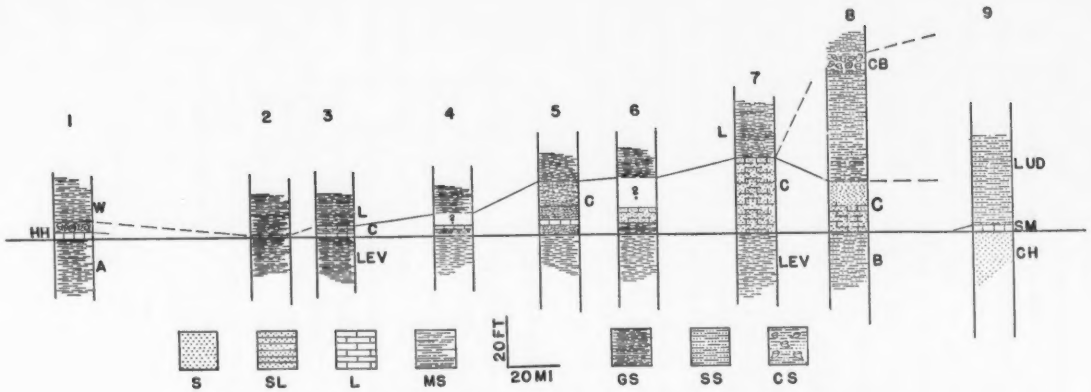

FIGURE 2.—*Diagram showing Centerfield formation in New York and Ontario*

ROCK TYPES

(CS) corals in shale and sandstone
(GS) dark-gray shale
(L) limestone
(MS) marly shale
(S) sandstone
(SL) sandy limestone
(SS) sandy shale or shaly sandstone

FORMATIONS

(A) Arkona shale
(B) Butternut shale
(C) Centerfield formation
(CB) Coral bed
(CH) Chenango sandstone
(HH) Hungry Hollow formation
(L) Ledyard shale
(Lev) Levanna shale
(Lud) Ludlowville sandy shale
(SM) Stone Mill limestone
(W) Widder shale

LOCALITIES

(1) Hungry Hollow, 2½ miles east of Arkona, Ontario.
(2) Lake Erie, New York.
(3) Blossom, Erie County, New York.
(4) 3 miles west of East Bethany, New York.
(5) Livonia Salt Shaft, Livonia, New York.
(6) One mile north of Centerfield, New York.
(7) Deans Creek, Cayuga Lake, New York.
(8) Skaneateles Lake, New York.
(9) 4 miles south of Hamilton, Chenango Valley, New York.

Tichenor, and *Vitulina* extends upward to the top of the Hamilton. In each occurrence these important fossils always appear with other Hamilton species that are characteristic of a level. In the second instance *Spirifer venustus* of western New York is much more robust than in Michigan and north-central Indiana. Specimens from Kentucky and southern Indiana on the other hand are more like the New York form.

CENTERFIELD FORMATION OF NEW YORK

At its type section on Schaffer Creek, about a mile north of Centerfield, Canandaigua quadrangle, the Centerfield consists of 5 feet of alternating shale and limestone bands, the latter up to 4 inches thick. This is overlain by about 6 feet of sandy limestone abounding in corals. The formation thins westward to 4½ feet at Blossom in Erie County where it is mostly limestone. West of Blossom on the Lake Erie shore the Centerfield has not been seen, but its approximate position is marked by a thin layer of limestone containing abundant *Styliolina* and called the "Pteropod bed" by Grabau.

East of the type section the sand content of the Centerfield increases, and the differentiation into alternate bands of shale and limestone disappears. On Seneca and Cayuga lakes it is a fairly homogeneous sandy limestone about 30 feet thick, but at Skaneateles Lake considerable cross-bedded sandstone appears. Furthermore, at this place the Centerfield fauna appears also in the overlying dark, soft sandy shale forming the lower part of the Ludlowville. The coral bed about 50 feet above the top of the sandy part of the Centerfield at Fellows Falls and the "Cascades" on the east side of Onondaga Valley and in Butternut Valley may belong to the Centerfield. At any rate along the meridian of Skaneateles lake the Centerfield fauna occurs in a cross-bedded sandstone and sandy limestone and also in some few feet of the overlying dark sandy shale. Essentially this same arrangement continues as far east as the Chenango Valley. Across the Cazenovia quadrangle the Centerfield was found at three localities only. The easternmost manifestation of the Centerfield occurs in the glens on both sides of Chenango Valley and at the southeast end of Bradley Brook reservoir in the Morrisville quadrangle where a layer of crinoidal limestone and shell debris 3 feet thick and an undetermined thickness of overlying shaly sandstone contain the Centerfield fauna. The limestone has been named the Stone Mill member (Cooper and Warthin, 1941) as it forms pronounced waterfalls in all the glens in which it occurs. This member contains *Vitulina, Spirifer venustus,* and an undescribed species of *Prismatophyllum* having exceptionally large corallites. The shaly sand-

stone overlying the Stone Mill member abounds in fossils including *S. venustus* and many other common Hamilton types. *Spirifer venustus* occurs only in the lower part of the sandstone, but other Centerfield types are found as much as 20 feet above the limestone. The Centerfield and its fauna have not been certainly identified east of Chenango Valley.

In western New York the Centerfield fauna is best developed at the well-known locality along the D. L. and W. Railroad about 3 miles west of East Bethany where most of the characteristic species have been taken. *Centronella* and *Vitulina* are abundant in the Finger Lakes area, and *Spirifer venustus* occurs throughout the extent of the formation. *Pentagonia* has been found at the type locality and also at Paines Creek on the east side of Cayuga Lake. Many of the common specimens identified as *Meristella haskinsi* actually belong to the genus *Camarospira*. Corals become less numerous east of Canandaigua lake and, except for the Skaneateles Lake occurrences, are quite rare in the eastern third of the formation.

CENTERFIELD FORMATION OF PENNSYLVANIA

The Centerfield formation is well exposed in eastern Pennsylvania along State highway 90 up the west side of Brodhead Creek valley about 2 miles north of Stroudsburg. Here 23 feet of bluish calcareous sandstone contains a few corals in the lower 6 feet, a profusion of corals in the next 10 to 12 feet, and a few corals in the upper 5 feet. *Spirifer venustus* is common, and *Parazyga* and *Trematospira* occur with many of the other common Centerfield species. *Pentagonia* occurs north of Stroudsburg and also on the Buck Farm, 6 miles east of Leighton. According to Willard (1936) this coral fauna extends west from Pike county through Monroe, Carbon, Schuylkill, Lebanon, and Dauphin counties to Perry County. The locality north of Stroudsburg is lithologically most like the Centerfield of the Finger Lakes area.

HUNGRY HOLLOW FORMATION OF ONTARIO

In southwestern Ontario Cooper and Warthin (1941) renamed the Encrinal limestone and coral bed at the base of the Widder shale the Hungry Hollow formation with the type section in Hungry Hollow (Bartletts or Marshs Mill) on the Aux Sables River, 2½ miles east of Arkona. The formation consists of 2½ feet of grayish-brown crinoidal limestone overlain by 3½ feet of light-gray argillaceous limestone packed with corals. This same sequence occurs along the Aux Sables River at a number of localities and in the village of Thedford. The Crinoidal or Encrinal limestone of older reports contain many Centerfield species

such as *Pentagonia bisulcata, Spirifer venustus,* and *Camarospira.* The fauna of the coral bed is almost an exact duplicate of that of the New York Centerfield and is known to collectors all over the world. Along the Aux Sables River at "Number Four Hill" *Prismatophyllum* is common as float along the bank. The probable source of these specimens is the "coral bed" according to Charles Southworth, but the writers have not seen them in place.

FOUR MILE DAM FORMATION OF MICHIGAN

Cooper and Warthin (1941) named this formation from its type section under the Four Mile Dam on Thunder Bay River about 4 miles above Alpena. The limestone belongs to a reef of undetermined thickness overlain by the Norway Point formation and consists of broken corals, corals and shell debris with occasional bluish-green shale pockets. Beautifully preserved fossils are abundant; molluscs are common, but the brachiopods, corals, and echinoderms point to a Centerfield age. *"Strombodes" alpenensis* occurs with many other corals. *Spirifer venustus* is abundant but is smaller than the New York form. Other genera of significance are: *Camarospira, Camerophoria, Trematospira, Parazyga, Cyclorhina,* and abundance of *Dolatocrinus* and *Megistocrinus.*

In the southeast corner of the Thunder Bay Quarries Company quarry on the east side of Alpena, the *Spirifer venustus* fauna occurs in a gray, fine-grained limestone forming the basal 2½ feet of the Four Mile Dam formation and is underlain by the Dock Street Clay. A few of the species were also taken in the 6 feet of similar limestone above that containing *Spirifer venustus.* Many Centerfield elements also occur in the Alpena limestone and Dock Street clay.

ALPENA LIMESTONE AND DOCK STREET CLAY OF MICHIGAN

The Alpena limestone of VerWiebe (1926) included more than 120 feet of limestone containing several distinct units. Warthin and Cooper redefine the Alpena to exclude the Genshaw and Newton Creek from the base and the Four Mile Dam limestone from the top. This leaves as a restricted Alpena about 79 feet of light-gray to white limestone containing a 1-foot bed of blue argillaceous limestone 57 feet below the top. This interval contains the large *Prismatophyllum* and *"Stromatopora"* reefs for which the Alpena region has long been noted.

The fauna of the Alpena is very difficult to place. The abundant corals, including *Alveolites* and *Favosites turbinatus* which are rare, suggest relationship to the Centerfield. The Dock Street clay, a part of the Alpena, shows still stronger affinities with the Centerfield.

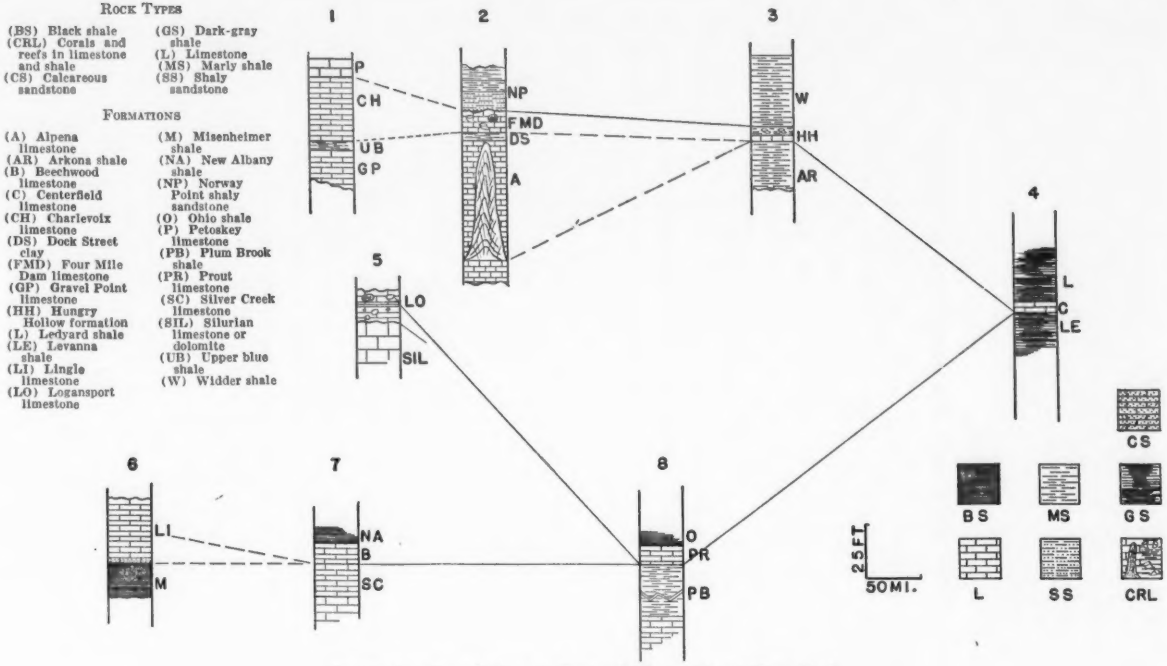

FIGURE 3.—*Correlation of Centerfield outside of New York*

LOCALITIES
(1) Emmet and Charlevoix counties, Michigan.
(2) Alpena County, Michigan.
(3) 2½ miles east of Arkona, Ontario.
(4) Vicinity of East Bethany, New York.
(5) Pipe Creek Falls, east of Logansport, Indiana.
(6) Branch of Lingle Creek, Union County, Illinois.
(7) Clark County, Indiana.
(8) Vicinity of Prout, Erie County, Ohio.

The Dock Street clay, a large lens of blue calcareous shale at the top of the Alpena, was best exposed during the stripping operations for the quarry of the Thunder Bay Quarries Company. Test drilling between the quarry and the lake shore revealed no clay at the level of the Dock Street. The lens is therefore too local to deserve a formal name. The Dock Street abounds in fossils and contains specimens of *Megistocrinus* and *Dolatocrinus* resembling those from the Beechwood of southern Indiana. The brachiopods *Camarospira, Camarotoechia thedfordensis, Cyclorhina, and Trematospira,* although rare, suggest relationship with the Centerfield.

The writers believe that the faunal content of the Alpena, Dock Street, and Four Mile Dam indicates their correlation with the Centerfield. The fauna of the Four Mile Dam is most like that of the Encrinal bed of the Hungry Hollow formation, while the fauna of the Alpena proper is more like that of the coral bed overlying the Encrinal. These apparent inversions are perplexing, but, inasmuch as the fossils from the base of the Alpena reefs through the Four Mile Dam limestone are Centerfield types, it seems best to correlate the interval with the Centerfield although such a correlation may not be precise.

GRAVEL POINT LIMESTONE OF MICHIGAN

This brittle gray limestone, over 100 feet thick, contains three shaly bands. One thin shale at the base contains *Chonetes emmetensis;* a second layer, the "lower blue shale" of Pohl (1930), contains many corals; and the third or "upper blue shale" caps the formation. The lowest shale contains fossils suggestive of those of the lower Alpena, while the "upper blue shale" has yielded species suggestive of the Dock Street clay. The Gravel Point is therefore correlated with the Alpena limestone and tentatively assigned to the Centerfield.

CHARLEVOIX LIMESTONE OF MICHIGAN

This limestone lies above the Gravel Point and beneath the Petoskey limestone. The "upper blue shale" of the Gravel Point indicates Dock Street affinities; fossils at the base of the Petoskey in the Curtis Quarry at Bayshore suggest the Norway Point. If these suggestions be correct, the Charlevoix limestone occupies the position of the Four Mile Dam limestone. Few fossils indicating any relationship to the Four Mile Dam limestone have been taken from the Charlevoix, and no faunal basis for correlation exists. The Charlevoix contains oölites, limestone breccia, and thin platy, mud-cracked limestone all suggesting near-

shore deposition. It is therefore suggested that the Charlevoix represents a possible westward shore phase of the Centerfield.

PROUT LIMESTONE OF OHIO

Overlying the Plum Brook shale and overlain by the black Ohio shale, 8 or 9 feet of hard limestone abounds in corals and other poorly preserved fossils. Dr. Erwin Stumm of Oberlin College has found the following species that date the Prout as Centerfield: *Pentagonia bisulcata, Spirifer venustus,* and *Camarospira.*

TENMILE CREEK DOLOMITE OF OHIO

This formation consists of about 11 feet of cherty limestone and dolomite containing many corals and other fossils. The writers collected *Dolatocrinus, Megistocrinus,* and *Cyclorhina.* Stauffer (1909, p. 147) reports *"Strombodes" alpenensis.* These fossils are believed to be of Centerfield age. Inasmuch as the Silica shale underlies the Tenmile Creek this assignment definitely places it between the Centerfield and the underlying Dundee.

LOGANSPORT LIMESTONE OF INDIANA

This formation is composed of about 12 feet of limestone, the lower half containing mostly corals and stromatoporoids, but the upper half abounds in a variety of fossils. The type section is at Pipe Creek Falls about 2 miles above the junction with the Wabash River, 7 miles above Logansport. In addition to corals the upper half contains *Megistocrinus, Dolatocrinus, Camarospira, Camerophoria, Cyclorhina, Parazyga,* and *Spirifer venustus. Eunella attenuata,* formerly known only from the Hungry Hollow formation, also occurs here. The Logansport formation had hitherto been identified as Jeffersonville limestone (Kindle, 1900, p. 561-569) because of the supposed presence of *Spirifer acuminatus.* The writers collected specimens of this brachiopod and found them to be related to *Spirifer granulosus* rather than to the Onondaga species.

BEECHWOOD LIMESTONE OF SOUTHERN INDIANA AND NORTHERN KENTUCKY

Butts gave this name to a crinoidal limestone 3 to 10 feet thick that forms the top of the Middle Devonian sequence in Jefferson County, Kentucky and Clark County, Indiana. This limestone abounds in fossils including *Alveolites goldfussi, Favosites placentus, F. turbinatus, Megistocrinus, Dolatocrinus, Eleutherocrinus cassedayi, Centronella impressa, Vitulina pustulosa, Spirifer sculptilis, S. venustus, Camarospira, Camero-*

phoria gainesi, Pentagonia bisulcata, Parazyga hirsuta, and *Cyclorhina nobilis.* The fauna of the Beechwood is almost identical with that of the Centerfield, and no more precise interregional correlation could be found. The Centerfield equivalent in this area forms the top of the Hamilton.

BOYLE LIMESTONE IN EAST-CENTRAL KENTUCKY

Foerste (1906) applied this name to the Middle Devonian limestone on the east side of the Cincinnati arch. He believed that it contained a mixture of Jeffersonville and Hamilton species. From his list it is clear that representatives of the Beechwood or Centerfield equivalent are likewise present. The following species, reported by Foerste, according to the writers' experience, occur in the Beechwood only: *Centronella glansfagea* = *C. impressa, Cyclorhina noblis* and *Pentagonia unisulcata* = *P. bisulcata.* These are thought to represent the Beechwood horizon in the Boyle.

LINGLE LIMESTONE OF SOUTHWESTERN ILLINOIS

The writers restrict this name to the limestone exposed above the Misenheimer shale on a branch of Lingle Creek in SW. ¼ sec. 26, T. 13 S., R. 2 W., about 3 miles west-northwest of Mill Creek, Illinois, where 31 feet of limestone is exposed in the slope above the dark shale. Some of the layers are crinoidal, while others contain considerable sand. A hard cherty layer slightly more than 2 feet thick and 16 feet above the shale contains *Tropidoleptus carinatus, Centronella impressa, Vitulina pustulosa, Spirifer sculptilis,* and scattered corals. The top exposed 15 feet of the formation contains fewer fossils, and the very top of the limestone is not known. The Lingle limestone at this place is correlated with the Beechwood limestone that also contains *Centronella* and *Vitulina.*

This Lingle limestone as exposed on the branch of Lingle Creek is not to be confused nor correlated with the Hamilton limestone above the Grand Tower limestone at the Devils Backbone and Bake-Oven north of Grand Tower, Illinois. This limestone is in part of Marcellus age and probably correlates with the St. Laurent limestone of the Little Saline Valley, Missouri, rather than with the Lingle described above. Neither *Centronella, Vitulina,* nor any other diagnostic Centerfield species has yet been taken from the St. Laurent. A possible correlate of the Lingle occurs at the top of the St. Laurent in a small creek just southeast of Union School, ¾ mile south of Seventy-six, Altenburg quadrangle. It is composed of crinoidal limestone and abounds in corals. Although diagnostic Centerfield species are not yet known from this limestone, the layer lithologically and in its abundance of corals suggests the Lingle.

CLIFTY LIMESTONE OF ARKANSAS

This limestone consists of 2½ feet of light-gray limestone exposed at the "head of a large ravine on the north side of the stream, on the west side of SW¼ sec. 17, T. 19 N., R. 27 W.", Eureka Springs quadrangle. Cooper re-examined the fossils collected from the Clifty by Purdue and Miser (1916) and identified *Centronella impressa* and *Vitulina pustulosa* as well as *Spirifer sculptilis* and the other fossils listed by Purdue and Miser. These species indicate a correlation with the Lingle limestone *(sensu stricto)* of southwestern Illinois.

PEGRAM LIMESTONE OF TENNESSEE

In Davidson and Cheatham counties Middle Devonian limestone and sandstone are exposed along the Harpeth River west of Nashville. Peoples (1931) described Onondaga and Hamilton faunas from the Pegram limestone that had hitherto been regarded as Onondaga only. Examination of Peoples' and Pohl's collections by Cooper showed no unequivocal Onondaga species. The forms recorded by Peoples as *Spirifer gregarius* are clearly misidentified. Of the two specimens one is an *Elytha* and the other an immature *Spirifer* of uncertain relationships. Part of the Pegram is certainly Hamilton, but the lower part heretofore assigned to the Jeffersonville may also be of Hamilton age. The presence of many corals in the upper Pegram and a few other forms known from the Beechwood, although not diagnostic Centerfield species, nevertheless indicate a possible correlation.

Pegram limestone is exposed in the west bluff of Mill Creek a mile north of Topsy, Waynesboro quadrangle, southern Tennessee. This section consists of about 6 feet of hard gray limestone. Hamilton fossils such as *Cypricardinia indenta* occur at about the middle of the mass. At the base a turbinate *Favosites* and a large wrinkled *Schuchertella* were found and are suggestive of Hamilton rather than Onondaga age. Possibly the Pegram correlates with the Sellersburg of Kindle rather than with the Jeffersonville and Sellersburg, but the point cannot be settled without more extensive collecting.

RAGLAND SANDSTONE OF ALABAMA

This formation in St. Clair County consists of 50 feet of fine-grained sandstone. Near the base is a fauna of equivocal age containing Marcellus and Centerfield elements. Of the latter *Spirifer venustus* occurs with *Douvillina* and *Elytha*. The Marcellus elements consist of *Coelospira* and *Leptaena*. The former is known at the base of the black shale on the east side of Cayuga Lake, New York, and *Leptaena* occurs in the

lower part of the Marcellus west of Catskill, New York, and in the Delaware and Dundee limestones of the Midwest. *Spirifer venustus* is known from the Marcellus south of Berne, Albany County, New York. Therefore, although the Ragland has some Centerfield affinities, it seems best to correlate it with the *Meristella*-coral zone 250 feet above the Onondaga south of Berne where *Spirifer venustus* is known.

AGE OF CENTERFIELD LIMESTONE

Clarke and Luther (1904), who defined this formation, placed it at the base of the Ludlowville shale. Recently Dr. Burnett Smith (1935) transferred the Centerfield from the Ludlowville to the top of the Skaneateles because of the transition between the two. The transitional nature of this contact is well exhibited from Skaneateles Lake eastward to Chenango Valley. The writers do not regard a transition as sufficient reason for transferring the formation from the Ludlowville.

Although the contact of the Skaneateles and Centerfield is transitional in east-central New York, this condition does not hold in western New York where the dark Skaneateles shale changes abruptly to light-gray calcareous shale and thin limestone beds. The Centerfield fauna is not transitional with the Skaneateles fauna below as it brings in a host of new types. The fossils from Skaneateles Lake to Chenango Valley extend into the overlying Ludlowville shales, thus tying the Centerfield to the upper beds.

STRATIGRAPHIC IMPORTANCE OF CENTERFIELD CORRELATIONS

The correlations demonstrated herein for the first time indicate the parts of the Hamilton represented by scattered outcrops in the Midwest and South. In Ontario the Hungry Hollow formation separating the Arkona and Widder shales indicates the Skaneateles age of the Arkona and the Ludlowville age of the Widder. On other grounds the age of the Ipperwash above the Widder can be demonstrated as Ludlowville. Insofar as known no Moscow exists on the Peninsula.

In Michigan the Four Mile Dam and Alpena limestones separate a thick Skaneateles sequence consisting of the Bell shale, Rockport Quarry limestone, Ferron Point shale, Genshaw formation, and Newton Creek limestone from the relatively thin Norway Point limestone of Ludlowville age. Higher beds occur in the Traverse, but they are assigned by Warthin and Cooper to the Tully or Geneseo. In Ohio identification of the Prout as Centerfield denotes a Skaneateles age for the Plum Brook shale.

In southern Indiana the Beechwood overlies the Silver Creek limestone; the latter overlies the Marcellus (Delaware) limestone containing

TABLE 1.—*Distribution of important Centerfield species*

Distribution of important Centerfield species	Stroudsburg, Pennsylvania	Chenango Valley, New York	Cayuga Lake, New York	Canandaigua Lake, New York	East Bethany, New York	Thedford and Arkona, Ontario	Prout limestone, north-central Ohio	10-Mile Creek limestone, northwest Ohio	Alpena County, Michigan	Pipe Creek Falls, Indiana	South-central Indiana, North-central Kentucky	Southwestern Illinois	Northwestern Arkansas
1. *Alveolites goldfussi* Billings				x	x	x			x		x		
2. *Favosites placentus* Rominger		x		x	x	x		x	x		x		
3. *F. turbinatus* Billings					x	x			x		x		
4. *Prismatophyllum* sp.		x			?	x			x	x			
5. "*Strombodes*" *alpenensis* Rominger								x	x		x		
6. *Dolatocrinus* sp.					x	x		x	x	x	x		
7. *Eleutherocrinus cassedayi* (Shumard and Yandell)					?	x					x		
8. *Megistocrinus* sp.				x	x	x		x	x	x	x		
9. *Camarospira* sp.				x	x	x	x		x	x	x		
10. *Camarotoechia thedfordensis* (Whiteaves)					x	x			x				
11. *Camerophoria* sp.				x	x	x	x		x	x	x		
12. *Centronella impressa* Hall											x	x	x
13. *Chonetes coronatus* (Conrad)	x	x	x	x	x	x		x			x		
14. *Cyclorhina nobilis* (Hall)					x	x		x	x	x	x		
15. *Douvillina inaequistriata* (Conrad)	x	x	x	x	x	x					x		
16. *Elytha fimbriata* (Conrad)	x	x	x	x	x	x			x	x	x	x	
17. *Eunella attenuata* (Whiteaves)						x				x			
18. *Nucleospira concinna* (Hall)	x	x	x	x	x	x		x			x		
19. *Pentagonia bisulcata* Hall	x		x	x		x	x				x		
20. *Spirifer venustus* Hall	x	x	x	x	x	x	x	x	x	x	x		
21. *S. sculptilis* Hall	x	x	x	x	x	x					x	x	x
22. *Tropidoleptus carinatus* (Conrad)	x	x	x	x	x	x		x			x		x
23. *Vitulina pustulosa* Hall	x	x	x								x	x	x
24. *Parazyga hirsuta* (Hall)	x			x	x	x			x	x	x		
25. *Richina truncata* Coryell and Malkin					x	x			x				

the coral *Hadrophyllum dorbignyi*. Thus the Silver Creek is of Skaneateles age. In southwestern Illinois the Lingle limestone overlies the Misenheimer shale probably of Skaneateles age. In south-central Missouri a possible correlate of the Lingle was indicated on a creek south of Seventysix. As this limestone overlies the St. Laurent of this area a Skaneateles age for that limestone is suggested, and this view is supported by the fossils of the St. Laurent.

REFERENCES CITED

Butts, C. (1915) *Geology and mineral resources of Jefferson County, Kentucky,* Ky. Geol. Survey, 4th ser., vol. 3, pt. 2, p. 120-129.

——— (1927) *The Devonian of Alabama,* Am. Jour. Sci., 5th ser., vol. 14, p. 373.

Carman, J. E. (1935) *A study of the Lucas County, Ohio-Monroe County, Michigan monocline and stratigraphy of northwestern Ohio and southeastern Michigan,* Guide 5th Ann. Field trip, Sect. Geol. Mineral., Mich. Acad. Sci., May.

Clarke, J. M., and D. D. Luther (1904) *Stratigraphic and paleontologic map of Canandaigua and Naples Quadrangles,* N. Y. State Mus. Bull. 63, p. 17-20, 47, 48.

Cleland, H. F. (1903) *Fauna of the Hamilton formation of the Cayuga Lake section in central New York,* U. S. Geol. Survey, Bull. 206, p. 20, 23, 24.

Cooper, G. A. (1930) *Stratigraphy of the Hamilton group of New York,* Am. Jour. Sci., 5th ser., vol. 19, p. 223.

——— (1933) *Stratigraphy of the Hamilton group of eastern New York,* Am. Jour. Sci., 5th ser., vol. 26, p. 548-549.

——— (1941) *New Middle Devonian stratigraphic names,* Wash. Acad. Sci., Jour., vol. 31, no. 6, p. 259-260.

——— and Warthin, A. S., Jr. (1936) *New correlations of Hamilton rocks,* (Abstract) Geol. Soc. Am., Pr., 1935, p. 376.

Foerste, A. F. (1906) *The Silurian, Devonian, and Irvine formations of east-central Kentucky, with an account of their clays and limestones,* Ky. Geol. Survey, Bull. 7, p. 10, 92.

Grabau, A. W. (1898) *Geology and paleontology of Eighteenmile Creek and the lake-shore sections of Erie County, New York,* Buffalo Soc. Nat. Hist., Bull. 6.

——— (1899) *The faunas of the Hamilton group of Eighteenmile Creek and vicinity in western New York,* N. Y. State Geol., 16th Ann. Rept., p. 291.

——— (1917) *Age and stratigraphic relations of the Olentangy shale of central Ohio, with remarks on the Prout limestone and so-called Olentangy shales of northern Ohio,* Jour. Geol., vol. 25, no. 4, p. 337-343.

Kindle, E. M. (1900) *The Devonian fossils and stratigraphy of Indiana,* Ind. Dept. Geol. Nat. Res., 25th Ann. Rept., p. 530-570.

Peoples, J. W. (1931) *Stratigraphy of the Middle Devonian of the Tennessee Basin,* Ill. Acad. Sci., Tr., vol. 23, no. 3, p. 431-439.

Pohl, E. R. (1930) *The Middle Devonian Traverse group of rocks in Michigan,* U. S. Nat. Mus., Pr., vol. 76, art. 14, p. 1-34.

Purdue, A. H., and Miser, H. D. (1916) *Eureka Springs-Harrison Folio, Arkansas-Missouri,* Geol. Atlas U. S. 202, p. 9.

Savage, T. E. (1920) *The Devonian formations of Illinois,* Am. Jour. Sci., 4th ser., vol. 49, p. 170, 176.

——— (1931) *The Paleontology of Kentucky,* Ky. Geol. Survey, 6th ser., vol. 36, *Devonian fauna,* p. 229-234.

Slocum, A. W. (1906) *A list of Devonian fossils collected in western New York with notes on their stratigraphic distribution,* Field Columbian Mus., Geol. ser., vol. 2, no. 8, p. 257-265.

Smith, B. (1935) *Geology and mineral resources of the Skaneateles Quadrangle,* N. Y. State Mus., Bull. 300, p. 34-44, 104-105.

Stauffer, C. R. (1909) *The Middle Devonian of Ohio,* Ohio Geol. Survey, 4th ser., Bull. 10, p. 119-122, 144-149.

Stauffer, C. R. (1915) *The Devonian of southwestern Ontario,* Canada Dept. Mines, Geol. Survey, Mem. 34, p. 156-180.

Stewart, G. A. (1938) *Middle Devonian corals of Ohio,* Geol. Soc. Am., Spec. Pap. 8, p. 7.

VerWiebe, W. A. (1926) *The stratigraphy of Alpena County, Michigan,* Mich. Acad. Sci., Pap., vol. 7, p. 181-192.

Warthin, A. S., Jr., and Cooper, G. A. (1935) *New formation names in the Michigan Devonian,* Wash. Acad. Sci., Jour., vol. 25, no. 12, p. 524-526.

Whitlatch, G. I., and Huddle, J. W. (1932) *The stratigraphy and structure of a Devonian limestone area in Clark County, Indiana,* Ind. Acad. Sci., Pr., vol. 41, p. 363-390.

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INTRUSIVE DOMES OF THE CLAREMONT-NEWPORT AREA, NEW HAMPSHIRE

BY CARLETON A. CHAPMAN

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ABSTRACT

Study of the Croydon and Unity domes sheds considerable light on the mode of intrusion of the Oliverian magma series in western New Hampshire. Each dome con-

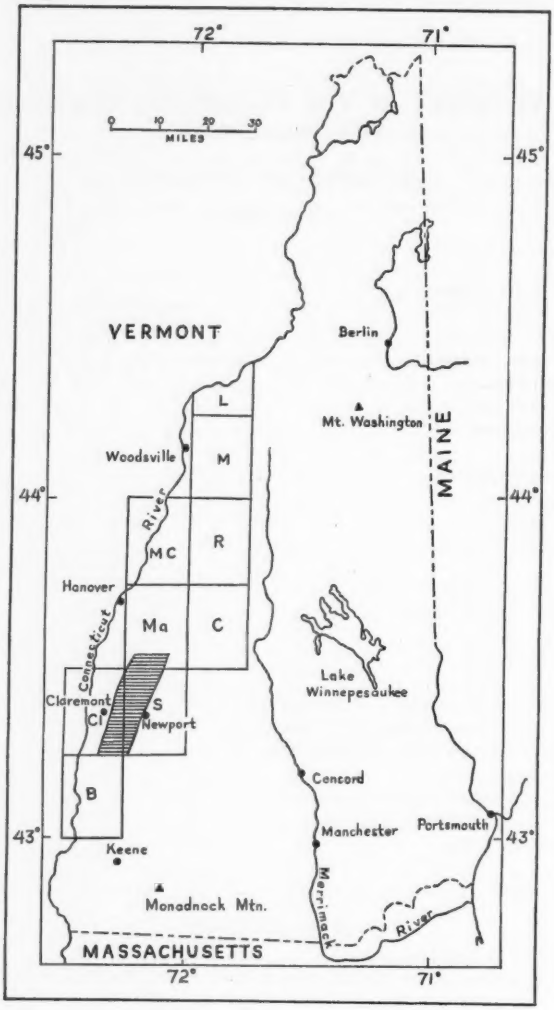

FIGURE 1.—*Index map of New Hampshire*

B=Bellows Falls quadrangle, C=Cardigan quadrangle, Cl=Claremont quadrangle, L=Littleton quadrangle, M=Moosilauke quadrangle, Ma=Mascoma quadrangle, MC=Mount Cube quadrangle, R=Rumney quadrangle, S=Sunapee quadrangle; Shaded area=Claremont-Newport area.

sists of an igneous core (late Devonian) surrounded by concentric belts of older (Ordovician?, Silurian, and early Devonian) metamorphosed sedimentary and volcanic rocks. The foliation of the plutonic cores is chiefly a primary flow structure as indicated by the parallel arrangement of inclusions, the parallelism between foliation and contacts, and the marked development of foliation near contacts. The bedding-plane schistosity of the surrounding metamorphic rocks is to be attributed to effects of the magma because (1) it is confined to the rocks immediately surrounding the igneous cores; (2) it parallels the contacts of the igneous cores; (3) it is relatively flat-dipping over large areas and thus rules out regional compression as a possible explanation. Several lines of evidence show that the numerous Oliverian domes are laccoliths: (1) The structure of each dome resembles that of a laccolith; (2) all the domes have roofs at the same horizon; (3) the intense folding and crumpling of the metamorphic rocks between domes indicate crowding of adjacent and isolated igneous cores; (4) laccolithic intrusion best explains the bedding-plane schistosity of the surrounding metamorphic rocks; (5) laccolithic intrusion best explains the foliation of the igneous cores. The eastern contact of the Mt. Clough pluton (New Hampshire magma series) is briefly considered, and the mode of intrusion of the body is explained.

INTRODUCTION

The Claremont-Newport area embraced in this study lies in western New Hampshire and includes parts of the Mascoma, Sunapee, and Claremont quadrangles (Fig. 1). It is within the New England Upland Section, a moderately dissected and glaciated upland developed on a complex rock structure. The relief averages only a few hundred feet. Rock structure is only locally reflected in the topography, although there is a fair correlation between topography and lithology. The highest elevations are held up by quartzite and quartz conglomerate as, for example, Croydon Peak (2781 feet), Grantham Mountain (2661 feet), Green Mountain (2005 feet), and Perry Mountain (1391 feet). The staurolite schist of Unity Mountain is also resistant. The igneous rocks, on the other hand, are much less resistant and underlie the more extensive lowlands. Much of the area, particularly the lowland, is covered with glacial and fluvio-glacial deposits which, in many places, make detailed mapping difficult.

The writer (Chapman, 1939) mapped the Mascoma quadrangle during the summer of 1936, but a brief review of the geology seems advisable here in order to give a complete picture of certain geological units included in the present study. The newly mapped area, in the Sunapee and Claremont quadrangles, required 8 weeks of field work, carried on principally during the summers of 1937 and 1938 and for short periods during 1939 and 1940.

Photostatic copies of the United States Geological Survey topographic sheets, enlarged to a scale of 3 inches to 1 mile, were used as base maps, on which all pertinent data were located by aid of aneroid barometer and pace and compass traverses.

The writer is especially indebted to his brother, Dr. Randolph W. Chapman of The Johns Hopkins University, for assistance in the field

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GENERAL GEOLOGICAL RELATIONS

Throughout most of the area large bodies of gneiss have invaded schist, quartzite, and amphibolite. Two plutonic masses of the Oliverian magma series have displaced the surrounding metamorphic rocks to form the Croydon dome in the northern part of the area and the Unity dome to the southwest (Pl. 1). The schistosity of the surrounding metamorphic rocks is parallel to the bedding and will be referred to as the bedding-plane schistosity. In general it dips outward from the plutonic masses. The igneous masses are believed to be laccoliths which domed and metamorphosed sedimentary and volcanic rocks and superimposed upon them the bedding-plane schistosity. As will be shown, the rocks of the Oliverian magmas were intruded before regional compression and metamorphism.

The extreme eastern part of the area is occupied by the Mt. Clough pluton, a huge sheetlike intrusion composed of Bethlehem gneiss, which developed a primary flow structure parallel to its low-dipping walls. The Bethlehem gneiss is believed (Billings, 1937, p. 538; Fowler-Lunn and Kingsley, 1937, p. 1381; Chapman, 1939, p. 164; Hadley, 1942, p. 166) to have been intruded just after the cumulation of the regional folding.

METAMORPHIC ROCKS

GENERAL STATEMENT

The numerous belts of metamorphic rocks have been traced southwest from the Mascoma quadrangle (Fig. 1) where they were correlated (Chapman, 1939) with equivalent formations in their type localities in the Littleton and Moosilauke quadrangles (Billings, 1937). So similar are the rocks considered in this report to those of the Mascoma quadrangle that only a brief description of them will be given here. (*See* Chapman, 1939, p. 132-149, for further details.)

The rock types, more or less in order of decreasing abundance, are schist, fine-grained gneiss, quartzite, amphibolite, and granulite.[1] They

[1] The term granulite is used in this paper as defined by Holmes (1928, p. 111): "A granulose metamorphic rock composed of even-sized interlocking granular minerals."

were derived from both sedimentary and volcanic materials, and nearly all belong to the middle-grade zone of metamorphism. Rocks of the high-grade zone and rocks believed to be transitional between the middle-grade and high-grade zones are confined to the area between the Croydon and Unity domes. The stratigraphic sequence of the various formations is as follows:

Littleton formation	Lower Devonian (Oriskany)
Fitch formation	Middle Silurian (Niagaran)
Clough formation	Middle or Lower Silurian
----------Unconformity----------	
Ammonoosuc volcanics	Upper Ordovician (?)

AMMONOOSUC VOLCANICS

The rocks of the Ammonoosuc volcanics are amphibolite (including some hornblende schist) and fine-grained biotite gneiss. Evidence in adjacent areas indicates that these rocks were derived from volcanic tuff, breccia, and conglomerate. In the present area, however, practically all evidence of an original fragmental structure has been destroyed by metamorphism. It is quite possible that many true flows are represented in the section.

Amphibolite and hornblende schist are most abundant in the northern half of the area. They are dark green or black, medium-grained rocks with a schistose to granulose structure.[2] Many specimens show a definite linear arrangement of amphibole needles. Blue-green hornblende and andesine or labradorite, roughly in equal amounts, are the essential constituents. The accessories include epidote, magnetite, sphene, and apatite. Biotite and chlorite are rare. The mineral composition indicates that these rocks were probably originally basaltic.

Biotite gneiss constitutes most of the formation in the southern half of the area. It is light gray or buff and is medium- to fine-grained. The structure is most commonly gneissic or granulose but may be schistose if micas are abundant. Major constituents include quartz, oligoclase or andesine, and biotite. In some specimens muscovite is abundant, and in others garnet occurs in small crystals. Apatite, sphene, and epidote occur in very minor quantities. Chlorite, if present, is secondary, having altered from biotite or garnet. Originally these rocks probably had the composition of dacite. Soda rhyolite, so abundant in the Littleton and Moosilauke area (Billings, 1937, p. 475), is not common here. It is possible that the muscovite in some of the gneiss was derived from potash feldspar; such rocks would have originally been quartz latite.

[2] The term granulose structure is used in this paper as defined by Holmes (1928, p. 112): "A structure characteristically developed in granulites, due to the presence of granular minerals. . ."

The total thickness of the Ammonoosuc volcanics farther north is known to be 2000 feet (Billings, 1937, p. 12). In the Claremont-Newport area only a few hundred feet of volcanics is exposed below the overlying Clough formation. The volcanics are underlain by rocks of the Oliverian magma series which, as will be shown, are intrusive into them. The most reliable measurements of thickness were obtained on the east slope of Croydon Mountain where the structure of the rocks is comparatively simple. Here 300 feet of the formation is exposed.

CLOUGH FORMATION

Quartzite, quartz conglomerate, mica schist, staurolite schist, and sillimanite schist constitute the Clough formation. The quartzite is medium-grained to fine- grained and is composed essentially of granoblastic quartz with a little biotite, muscovite, and plagioclase. It is generally white or gray, although limonite staining is common. Quartzite grades into, or is interbedded with, quartz conglomerate. The conglomerate contains large flattened pebbles of vein quartz and quartzite embedded in a white to buff matrix. Quartzite is an abundant rock type, but quartz conglomerate is uncommon and is found principally on Croydon Mountain.

Mica schist constitutes a large proportion of the formation and occurs throughout the area. The schist is relatively coarse-grained and commonly finely crinkled. Quartz and muscovite are the essential minerals, but locally biotite, garnet, and oligoclase are abundant. With an increase in percentage of quartz, mica schist grades through mica-quartz schist and micaceous quartzite into true quartzite.

Staurolite schist, not previously described from the Clough formation, occurs on Green Mountain and locally south of Unity Mountain. It contains porphyroblasts of staurolite up to 2 inches long. In most places the porphyroblasts have altered to muscovite and chlorite, but rough outlines of the original crystals still remain. The presence of staurolite in the schist of the Clough formation indicates an original sediment relatively rich in alumina.

Another alumina-rich rock is sillimanite schist abundant in the Clough formation between the Croydon and Unity domes. This rock is characterized by a coarse texture and a schistose to gneissic structure typical of high-grade metamorphism. Much of the sillimanite is in small knots, usually less than an inch long, which resist weathering and stand out on weathered surfaces. Microscopic study shows sillimanite abundant in stringers and bundles oriented more or less parallel to the foliation and replacing layers of biotite. Quartz, biotite, and muscovite in addition to sillimanite are the essential minerals.

Amphibolite sills are common in the Clough formation. One of the largest, at least 10 feet thick, is well exposed ⅔ mile northwest of Unity. The amphibolite is a medium-grained rock composed essentially of hornblende and andesine and is believed to represent metamorphosed basaltic material.

The maximum thickness of the Clough formation in the Mascoma quadrangle, immediately to the northeast, was estimated to be 1200 feet (Chapman, 1930, p. 139). More recent studies in the present area suggest that this estimate is perhaps too high and that the maximum for both the Mascoma quadrangle and Claremont-Newport area should be placed at about 800 feet. Sections across Croydon and Perry Mountains show that folding will explain the great breadth of the Clough belt. The minimum thickness of the formation is about 300 feet, unless the formation pinches out beneath the drift and valley fill along Sugar River, just east of Claremont.

FITCH FORMATION

The Fitch formation is represented chiefly by lime-silicate rocks and red-brown biotite schist. Lime-silicate granulites are abundant and well exposed on Croydon Mountain where they make up most of the formation. They are usually fine-grained to medium-grained with a green and white blotched or banded appearance. The green blotches or layers are composed chiefly of actinolite, oligoclase, and microcline, whereas diopside, oligoclase, and microcline constitute the white portions. Such mineral assemblages suggest that these granulites were derived from arenaceous dolomites.

Medium-grained to fine-grained biotite schist is most abundant in the southern part of the area. The essential constituents are red-brown biotite, actinolite, and calcite with varying proportions of microcline, oligoclase, and quartz. Minerals less commonly encountered are hornblende, muscovite, garnet, epidote, and sphene. The wide variation in mineral composition of these schists suggests an equally wide variation in the original sediments, which were undoubtedly calcareous, dolomitic, and arenaceous shales. A few thin beds of impure marble are interbedded with the schist 1½ miles north of Unity.

The thinning of the Fitch formation in western New Hampshire from 400-700 feet in the Littleton and Moosilauke quadrangles to 0-100 feet in the Mascoma quadrangle has been explained as due to differences in types of sediments deposited simultaneously in different parts of the State (Chapman, 1939, p. 139). It is believed that, while rocks of the Fitch type (lime-silicate rocks) were being deposited to the north, rocks of the Clough facies (quartzite and schist) continued for some time to

be deposited to the south. This explanation will account not only for the thinning of the typical Fitch rocks but also for the corresponding thickening of the Clough facies toward the south. The boundary between the Fitch and Clough formations must be mapped on the basis of lithology. Where the Fitch is absent the Littleton formation rests directly on the Clough. The Fitch formation in this area is somewhat discontinuous, and its thickness is estimated to range from 0-200 feet.

LITTLETON FORMATION

Fine- to medium-grained schists constitute the Littleton formation. Mica schist and mica-quartz schist are the most abundant and widely distributed rock types and commonly carry numerous porphyroblasts of biotite and garnet, some of which distort the schistosity whereas others cut sharply across it. The essential minerals are quartz, muscovite, and partly chloritized biotite.

Staurolite schist is abundant on the west slope of Croydon Mountain where staurolite occurs as porphyroblasts up to 5 inches long. In most places, however, this mineral has partly or completely altered to a mixture of sericite and chlorite. Staurolite schist is also abundant on Unity Mountin and is found locally in the Littleton formation to the south and on Green Mountain, Perry Mountain, and Bible Hill. Save for the numerous large porphyroblasts, the staurolite schist closely resembles the mica schist of the Littleton formation. The parent rock of the staurolite schist was probably an alumina-rich sediment.

Staurolite-sillimanite schist is abundant in the Littleton formation immediately north and northwest of Blueberry Ledge. Superficially this schist resembles the staurolite schist already described, but closer inspection, particularly on weathered surfaces, reveals the presence of considerable sillimanite. Both minerals may be observed in the same hand specimen. The sillimanite is principally in irregular knots up to an inch long. The large size and euhedral form of the staurolite crystals of the schist indicate that they must have formed during progressive metamorphism and are not products of retrograde metamorphism from sillimanite and biotite. Many of the staurolite porphyroblasts are free from alteration and in hand specimen show no association with sillimanite. Staurolite and sillimanite occur within the same thin section and appear independent of each other. Staurolite was not observed in contact with sillimanite although the former has in many instances partially altered to sericite and chlorite or has been replaced by coarse muscovite. It is believed that the presence of sillimanite and staurolite in these rocks indicates a transition between the middle-grade and high-grade zones of metamorphism.

Sillimanite schist occurs in the Littleton formation just west of Mt. Tug and closely resembles the mica schists of the Littleton formation save for a few per cent of sillimanite.

The relative amounts of sillimanite and staurolite in the schists of the Clough and Littleton formations between the Croydon and Unity domes are so variable and exposures of the schists so few that the writer was unable to locate boundaries between metamorphic zones. In general, however, there appears to be a local increase in intensity of metamorphism between the two domes. The significance of this local change will be considered later.

The intense folding of the Littleton formation makes any estimate of the stratigraphic thickness highly inaccurate. The rough estimate of 4000 feet gives only the order of magnitude of the maximum thickness of the formation exposed.

PLUTONIC ROCKS

GENERAL STATEMENT

The intrusive igneous rocks are medium-grained to coarse-grained gneiss belonging to two separate igneous series. The older or Oliverian magma series is represented by the Croydon and Unity groups, and the younger or New Hampshire magma series is represented solely by the Bethlehem gneiss. The rocks of the Oliverian magma series are considered to be younger than early Devonian (Billings, 1937, p. 502; Chapman, 1939, p. 146; Hadley, 1942, p. 143) because they displaced the Littleton formation during intrusion. They are believed to be older than the period of folding and metamorphism because they were affected by the deformation. Some of the effects of this deformation will be considered in the following pages. The Oliverian magma series may be best considered of late Devonian age although the evidence is not conclusive. The most recent data (Chapman, 1941) indicate that the Bethlehem gneiss was intruded just after the culmination of the folding but while stresses still existed in the crust. This intrusion is also, perhaps, best considered to be of late Devonian age.

CROYDON GROUP

The term Croydon group refers to the plutonic rocks of the Oliverian magma series within the Croydon dome (east of Croydon Mountain). The group consists of fine- to medium-grained biotite gneiss, locally subporphyritic with large crystals of white or pink microcline, commonly a centimeter long. The composition of the gneiss ranges from quartz diorite to quartz monzonite although quartz diorite and granodiorite constitute most of the pluton. Quartz, oligoclase, microcline, and biotite

are the essential minerals, but epidote, muscovite, and magnetite are minor constituents. A little myrmekite was observed in the microcline-rich types. Hornblende occurs in small amounts and only in those rocks close to the contact of the Ammonoosuc volcanics or near inclusions of the volcanics. Its presence is believed due to contamination by the volcanics. Epidote is generally more abundant in the border phases. The granularity of the gneiss is somewhat finer and the foliation more perfectly developed near the contacts. The low content of biotite (average 6%) in many places makes an accurate determination of the foliation difficult.

UNITY GROUP

The term Unity group is here proposed for the plutonic rocks of the Oliverian magma series within the Unity dome, southwest of the Croydon dome. This group is so similar to the Croydon group that a detailed description is not necessary. Microcline is scarce so quartz diorite is the predominating type. Epidote is most abundant in the hornblende-bearing varieties but is widely distributed.

BETHLEHEM GNEISS

The Bethlehem gneiss, a member of the New Hampshire magma series, is a medium- to coarse-grained biotite gneiss ranging from granodiorite to granite. It is locally porphyritic with microcline phenocrysts 1 or 2 inches long. The essential minerals are quartz, microcline, andesine, biotite, and a little muscovite. The minor constituents include epidote, zircon, apatite, and allanite. Considerable myrmekite was observed in some thin sections. The foliation of the gneiss has been somewhat deformed. The thin layers of mica commonly show evidence of crinkling, and microscopic study indicates some granulation of larger crystals. Where the foliation is poorly developed the mica shows a definite linear arrangement. Pegmatites, abundant in the gneiss and surrounding country rock, are genetically related to the gneiss itself.

The Bethlehem gneiss is distinguished from the rocks of the Oliverian magma series by its coarser grain, larger microcline phenocrysts, more perfect gneissic structure, more calcic plagioclase, higher biotite content, and presence of considerable muscovite.

STRUCTURE

GENERAL STATEMENT

The structure of the Claremont-Newport area is in general similar to that of most of western New Hampshire. It embraces features characteristic of folded, faulted, and metamorphosed sediments and volcanics as well as those typical of many plutonic intrusives. The major struc-

tural features, which occupy the greater part of the area, are the two Oliverian domes referred to in this report as the Croydon and Unity domes. Of somewhat less importance here is the Mt. Clough pluton which extends along the eastern edge of the area.

TYPICAL OLIVERIAN DOME

The Oliverian domes of western New Hampshire include a series of laccolithic bodies arranged in a narrow belt extending northeast-southwest, parallel to the trend of the major folds, for over 125 miles. Eight of these domes have already been studied, and others undoubtedly will be reported on as field mapping is continued farther to the north and to the south. The typical Oliverian dome consists of a central, igneous core of biotite gneiss surrounded by concentric belts of Ammonoosuc volcanics and successively younger formations. These surrounding rocks possess a good schistosity parallel to the bedding which is believed to have originally dipped outward from the plutonic core. Later regional compression, however, is believed to have folded these rocks to a greater or less degree so that the attitude of the bedding and schistosity was in many places considerably changed. The foliation of the intrusive near its margins is parallel to the contact and also to the bedding and schistosity of the surrounding metamorphic rocks. The foliation well within the plutonic mass, however, is generally poor and somewhat irregular.

CROYDON DOME

The Croydon dome lies mainly in the Mascoma and Sunapee quadrangles (Fig. 1) and extends north and south for about 12 miles (Pl. 1). The eastern part of the dome has been displaced by the Grantham fault making it impossible to determine the east-west dimension. It is very probable, however, that the original shape of the dome in ground plan was elliptical with the longest dimension trending slightly east of north.

Exposures of the Croydon group are poor, except near contacts, owing to extensive glacial drift. Foliation of the gneiss is best developed near the western and northern borders of the pluton, where it is parallel to the bedding and schistosity of the overlying Ammonoosuc volcanics. Toward the interior of the dome the foliation is poorly developed, and its direction and amount of dip are variable (Pl. 1). Near the borders of the pluton the gneiss is finer-grained, contains more mafic minerals (biotite), and is nearly always quartz diorite.

The contact between the Croydon group and Ammonoosuc volcanics is best exposed along the east slope of Croydon Mountain and at the north end of the dome. The relationship between the two rocks is well

shown near the "r" in "Croydon Peak" in Plate 1 (elevation 2480 feet on spur east from Croydon Peak). Here numerous small sills of quartz diorite gneiss cut hornblende and biotite rocks of the Ammonoosuc volcanics. Farther east (stratigraphically below) the sills of gneiss become larger or more numerous so that the relative amount of Ammonoosuc volcanics diminishes and the volcanics appear as slablike inclusions many feet long in the gneiss. The contact, therefore, is gradational, and the zone of transition 100 or 200 feet wide. Such is the nature of the contact wherever observed in the Croydon dome. The transition zone is so narrow and in many places so difficult to delimit that the contact has been shown as a single line in Plate 1.

A coarsely banded, transition rock was thus produced by the intimate association of the Oliverian magma and the Ammonoosuc volcanics. The magma was injected in thin sills along bedding planes of the volcanics, and the sills in places may be so abundant that the intersill rock now forms but a series of thin sheetlike inclusions oriented parallel to foliation of the intrusive quartz diorite. Thus bedding and schistosity of the volcanics parallel the foliation of the intrusive. Sills and intersill rock vary in thickness from a fraction of an inch to many feet. In most cases the contacts of the sills are gradational between the light-colored intrusive and dark-colored inclusion.

Reaction between inclusions and magma seems evident. In the smaller inclusions and in the thinner layers of intersill rock, biotite instead of hornblende has formed as the chief mafic mineral. Hornblendic material was probably made over into biotite, and this transformation may have robbed the magma of potash, leaving the sill material poor in microcline. This may explain why quartz diorite is generally confined to the margins of the pluton. In such a reaction epidote may also have been formed. Field and laboratory studies indicate that the content of epidote in the Croydon group increases toward the contact zone. This same relationship was first noted for the Mascoma dome in the Mascoma quadrangle (Fig. 1). Even the gneiss surrounding many large inclusions of amphibolite contains more epidote than that farther away. These relationships indicate that the original magma was locally somewhat contaminated by the included Ammonoosuc volcanics. Contamination was probably not confined to the contact zone but may have extended 1000 to 2000 feet into the pluton.

At the north end of the Croydon dome bedding and schistosity of the metamorphic rocks are parallel and, in general, dip to the north. Crumpling and open folding of these structures may be observed in most outcrops. Folds in bedding and schistosity range from less than an inch to many feet across and plunge 20° to 50° N. Somewhat larger folds,

hundreds of feet across, are indicated by the sinuous contact of the Ammonoosuc volcanics and the Clough formation (Pl. 1). To a less degree the winding contact between the Ammonoosuc volcanics and the gneiss of the Croydon group indicates a folded structure.

A traverse, at the northern border of the map, from the Grantham fault westward to the Littleton formation, shows a marked increase in intensity of folding. In general, folds become more numerous and more tightly formed toward the west.

Numerous traverses across that part of Croydon Mountain which lies north of Croydon Peak, from the gneiss of the Croydon group to the Littleton formation, also indicate a marked increase in intensity of folding to the west. Here, also, bedding parallels schistosity. In the Ammonoosuc volcanics extending northward from Croydon Peak, bedding and schistosity are nearly flat or dip gently to the west and northwest. Folds here are uncommon but where observed are generally broad and open. Bedding and schistosity of the Clough formation north of Croydon Peak, however, show more intense folding. Large folds are suggested by the opposed dips, some of which are shown in Plate 1. Smaller folds, up to many feet across, may be observed in most outcrops along Croydon Mountain. In general the axes of these smaller folds, and perhaps those of the larger folds as well, are essentially horizontal or plunge at low angles to the north. About a mile west of Grantham Mountain an anticline brings up a patch of quartzite of the Clough formation in the midst of the Littleton formation; apparently the Fitch formation here is very thin or absent. The great breadth of the Clough formation north of Croydon Peak is probably due, therefore, to folding of an originally gently dipping belt of rocks (Pl. 1, cross section AA′).

Bedding parallels schistosity in the belts of metamorphic rock extending from Croydon Peak southward as far as the Grantham fault. Here, however, these structures dip more uniformly and at higher angles than farther north. Very few dips of less than 50 degrees were observed. Folds in bedding and schistosity do not appear to be so common, but where observed they are usually isoclinal. Small folds may be seen along an old road across Croydon Mountain at an elevation of 1600 feet on the east slope (half a mile east of the "C" in "Croydon Mountain" in Plate 1). Here is an excellent cliff exposure of quartzite and schist of the Clough formation. The cliff face gives an east-west cross section and shows several isoclinal folds a few feet across. The folds plunge 5° NNE., and their axial planes are vertical. It is believed that folds are just as common in the metamorphic rocks south of Croydon Peak as they are north of that point, but south of Croydon Peak they are more difficult to observe because they are isoclinal or tightly formed and have nearly

horizontal axes and steeply dipping axial planes. Most outcrop surfaces closely parallel fold axes making folds difficult to detect. Where the surface cuts the axes at a large angle, however, tightly formed or isoclinal folds may be observed.

The eastern side of the dome is bounded by the Grantham fault, which throughout most of its extent has brought the Bethlehem gneiss into contact with the Croydon group. A small portion of what appears to be the roof of the Croydon dome is still preserved near Croydon Flat. Bedding and schistosity of the roof rocks here are parallel, and the gentle dips indicate very broad and open folds (Pl. 1). Flexures in the bedding and schistosity of the Ammonoosuc volcanics and Clough formations were observed in several localities, but folding is even less marked than at the north end of the dome.

UNITY DOME

The Unity dome lies en echelon to the Croydon dome, in the southeastern part of the Claremont quadrangle. Its longest axis is approximately 10 miles and trends a little to the east of north. The Unity dome, like the Croydon dome, has been displaced on the east by the Grantham fault. Erosion exposed two central areas of gneiss (Unity group) of the Oliverian magma series, one south of Green Mountain and the other east of Perry Mountain and Bible Hill. These areas of gneiss are now almost completely covered with glacial drift so that outcrops are scanty and little is known of the internal structure of the igneous core.

The contact or transition zone between the Unity group and the Ammonoosuc volcanics is very poorly exposed. It may be observed, however, about a mile due west of the "s" in "Glidden's Hill" and again about a mile southeast of the "e" in "Bible Hill" (Pl. 1). In general the features of this zone resemble those of the contact zone in the Croydon dome. The paucity of exposures and narrowness of the zone justify the use of a single line in Plate 1 to represent the contact between the Ammonoosuc volcanics and Unity group.

It is significant that quartz diorite is the predominating rock type of the Unity group. Field and laboratory studies of other Oliverian domes, particularly those which have been deeply eroded (Chapman, 1939, p. 166; Hadley, 1942, p. 137), indicate clearly that quartz diorite is more or less confined to a theoretical border zone, generally less than 2000 feet wide. The more microcline-rich types (granodiorite, quartz monzonite, and granite), however, constitute the central parts of the intrusives. Blue-green hornblende, like that of the Ammonoosuc volcanics, is more or less widely distributed through the rocks of the Unity group, and, as already noted, epidote is most abundant in the hornblende-bearing va-

rieties. In the plutonic cores of other domes (Chapman, 1939, p. 144; Hadley, 1942, p. 141), including the Croydon dome, hornblende is confined to these theoretical border zones or may occur in proximity to amphibolite inclusions well within the plutons. Furthermore, it has already been shown that the content of epidote generally increases toward the margins of the Oliverian domes.

The above facts suggest that those rocks of the Unity group, within a short distance of the surface, formed not far from the contact of the Ammonoosuc volcanics and are within a so-called border zone, near the roof of the pluton. In other words, a relatively large proportion of the igneous core is still covered by the Ammonoosuc volcanics, but erosion has stripped this cover in two areas and exposed the upper part of the Unity group. If the thickness of the border zone of the Unity group is comparable with that in other domes, then it might be concluded that the contact of the Ammonoosuc volcanics and Unity group, at its highest point above the two exposed areas of the igneous core, originally had a maximum elevation of about 2000 feet above the present erosion surface. This would indicate a gentle slope for the roof of the pluton, a feature apparently characteristic of all the Oliverian domes.

On the east side of the two igneous areas, bedding and schistosity in the Ammonoosuc volcanics dip away from the igneous contact at moderate or low angles, indicating an outwardly inclining dome roof as in the typical Oliverian dome. On the west side of the two igneous areas, however, it is a problem to explain the steep dips in bedding and schistosity of the volcanics if these volcanics also represent a relatively thin cover of the pluton. It is believed, nevertheless, that originally these volcanics dipped westward away from the axis of the dome, and that later they were tightly folded (in part isoclinally) thereby producing steep dips in bedding and schistosity. Evidence supporting this belief will now be given with a consideration of the Clough formation.

The continuous belt of the Clough formation extending west from Blueberry Ledge, across Green Mountain, and from there south to Bible Hill and Perry Mountain and east to the Grantham fault outlines the general shape of the Unity dome. Throughout this belt bedding and schistosity are parallel.

In the Clough formation on Green Mountain bedding and schistosity have been folded. Folds many feet across may be seen in quartzite and schist near an old triangulation station at the west end of the mountain (near the "G" in "Green Mtn." in Plate 1). Smaller folds were observed at many other places on the mountain, and these generally plunge 30°-50° NE. Large folds are suggested by the wavy contact of the Ammonoosuc volcanics and Clough formation.

The structure of the western flank of the Unity dome is believed to be similar to that of the western flank of the Croydon dome. From Green Mountain to Bible Hill the bedding and schistosity dip about 60° E., but farther south the dip steepens to 90° and then becomes reversed. Generally the westerly dips are above 70°, but locally values as low as 60° W. were recorded. A few steep easterly dips were noted along Perry Mountain. At the south end of the dome bedding and schistosity dip 15°-30° S.

Lenslike masses of schist occur at a number of places in the belt of Clough formation extending from Green Mountain to Perry Mountain. Four of these small patches of schist are shown in Plate 1. In most places they are staurolite schist petrographically identical with staurolite schist of the adjacent Littleton formation. These masses are interpreted, therefore, not as metamorphosed shaly lenses in the quartzite but as synclines of the Littleton formation with essentially horizontal axes, which were tightly folded down into the quartzite and preserved from erosion. The small patch of quartzite in the Littleton formation, just west of Bible Hill, is identical with that of the Clough formation. As no quartzite of this type is known in the Littleton formation, the small patch may be interpreted either as the exposed crest of an anticline which brings up the Clough formation from below or as a sliver of quartzite torn from the Clough beds and sheared up into the argillaceous rocks during intense folding. Regardless of whether the quartzite is an anticline or an isolated sliver, its presence strongly suggests intense folding just as do the small synclines of staurolite schist in the belt of Clough formation. The marked absence of opposed dips in bedding and schistosity in the Clough formation and adjacent rocks would indicate isoclinal folds. It is believed, therefore, that a series of nearly isoclinal folds with essentially horizontal axes and steep axial planes will explain the structure of the metamorphic rocks on the west flank of Unity dome (Pl. 1, cross sections CC′ and DD′).

South and east of Perry Mountain the structure of the metamorphic rocks is simpler. Here, at the south end of the dome, bedding and schistosity of the Clough formation and Ammonoosuc volcanics are parallel and dip southward at low angles. Only broad open folds are observed in outcrops.

The wide belt of the Clough formation between the Croydon and Unity domes shows intense folding, but bedding and schistosity are, nevertheless, parallel. A major syncline lies immediately to the southwest of Blueberry Ledge, and another, the central portion of which is occupied by the Fitch formation, lies immediately to the northeast. The dips in

bedding and schistosity range from 40° to 70° and are very consistently to the northeast. This, together with the great breadth of the Clough formation, strongly suggests tight or isoclinal folding.

GRANTHAM FAULT

The term Grantham fault is here proposed for a large gravity fault which was first recognized in the Mascoma quadrangle. In the course of the present work it was traced to its southern limit in the Claremont quadrangle. It has a total length of 31 miles. Field evidence in the Mascoma quadrangle indicates that this is a high-angle fault dipping east, with the eastern block dropped down. This high angle of dip as well as the relative displacement is probably maintained along the entire strike of the fault. For several miles north of Grantham the throw is calculated to be approximately 4000 feet, and the Bethlehem gneiss is in contact with rocks of the Oliverian magma series. The throw decreases to about 1000 feet northwest of Unity Mountain where the Littleton formation is in contact with the Ammonoosuc volcanics. From here southward the amount of displacement decreases gradually, and the fault terminates at the south end of Unity dome.

The fault was accurately located by numerous structural breaks and by brecciated and silicified zones along its course. The outstanding discordances of structure are at the north and south ends of Croydon dome and at the south end of Unity dome. The most prominent brecciated and silicified zones along the fault trace are shown in Plate 1. These may be easily observed in the field because they form highly resistant masses and usually cause small hills and ridges. At several places the gneiss of the Croydon group shows brecciation usually accompanied by intense silification and leaching of biotite. Three more or less gradational types of silicification are recognized in the gneiss: Quartz may fill fractures and openings between fragments; it may partially replace the gneiss in small grains making the rock more massive and extremely brittle; or, as on Glidden's Hill, it may occur in large blebs up to ¾ inch across resembling anhedral phenocrysts. The quartz blebs are usually almond-shaped with their longest dimension parallel to the plane of foliation.

METAMORPHIC ROCKS EAST OF GRANTHAM FAULT

In the metamorphic rocks east of the Grantham fault bedding and schistosity are parallel and are only slightly folded and flexed (Pl. 1, cross sections CC′ and DD′). A broad, open syncline lies immediately to the east of Glidden's Hill. On Glidden's Hill the underlying Clough formation comes to the surface on the west limb of the fold. A mile east of Glidden's Hill the Clough formation comes to the surface on the east

limb of the fold. A mile south of Glidden's Hill lime-silicate rocks of the Fitch formation are exposed on the east limb of the syncline. These beds, however, are discontinuous and pinch out to the north.

The Littleton schists have been broken and dropped down along a fault from a point a mile east of Glidden's Hill northeastward to just beyond Sugar River. This is believed to be a gravity fault although good field evidence supporting this conclusion is lacking. It seems probable, however, that the fault developed at essentially the same time as the Grantham fault and, therefore, under the same general conditions (probably tension). The staurolite schist of Unity Mountain occupies a graben which is bounded on the east by this fault and on the west by the Grantham fault (Pl. 1, cross section CC′). The vertical displacement along the west side of the graben probably does not exceed 1000 feet, and the maximum displacement along the east side is calculated to be 900 feet.

About 1½ miles east of Glidden's Hill an open syncline in the Clough formation has brought down beds of the Littleton and Fitch formation and preserved them as small isolated patches in the deeper parts of the fold. These isolated patches or outliers are linearly arranged parallel to the synclinal axis. The largest of these outliers is bounded on the east side by a small normal fault. A short distance east of these outliers the Clough formation forms an anticline, and still farther east it dips east under the Fitch and Littleton formations. As the narrow belt of Fitch here is traced southward it swings to the west and pinches out so that the Littleton formation rests directly upon the Clough formation. Here the contact swings around the southern end of the Grantham fault and then turns north toward Perry Mountain.

Extending north from the village of Unity to past Mt. Tug is another fault which throughout most of its extent separates Ammonoosuc volcanics on the east from the Clough and Littleton formations on the west. Pronounced effects of shearing and brecciation may be seen in the Bethlehem gneiss near the fault, a mile north of Unity. Similar effects, but to a less degree, were observed in the amphibolite of the Ammonoosuc volcanics near Sugar River, north of Mt. Tug. The maximum vertical displacement of the fault is calculated to be about 1000 feet.

MODE OF INTRUSION OF OLIVERIAN DOMES

The Oliverian domes have many general features which might lead one to interpret them as (1) laccoliths, (2) stocks or batholiths, (3) phacoliths, or (4) upwarps on a single folded sill. The writer prefers to consider the domes as laccoliths which intruded the Ammonoosuc volcanics at a level a few hundred feet below the base of the Clough formation. Evidence gathered from a study of several of these domes

seems definitely opposed to any of the other interpretations and more firmly substantiates the laccolithic theory.

The major objections to the last three theories may now be considered in detail. In view of recent studies it is more difficult to picture these

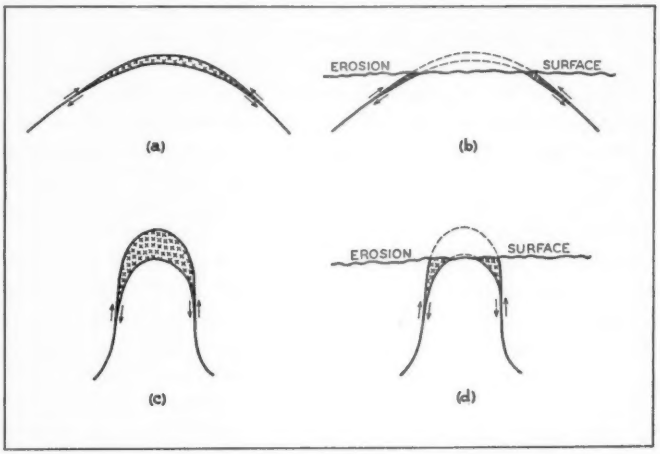

FIGURE 2.—*Effect of erosion on phacoliths formed in open and tightly closed folds*

(a) A broad arch in which slipping between beds, shown by arrows, is sufficient to form only a thin phacolith;
(b) the same phacolith nearly obliterated by erosion;
(c) a tightly formed arch in which slipping between beds has been sufficient to form a thick phacolith;
(d) the same phacolith with floor exposed only after considerable erosion.

domes as the tops of small batholiths or stocks, intruded either by stoping or forceful injection and cross-cutting the deeper rocks. Eight of these domes are known in western New Hampshire, and they are arranged in a narrow belt at least 125 miles long and a few miles wide. If they represent stocks or batholiths, why should they all have concordant roofs at essentially the same horizon? As far as it is possible to determine, the position of the roofs relative to the top of the overlying layer of volcanics does not vary more than a few hundred feet.

Two serious objections may be raised to a phacolithic theory of origin: (1) The arches of the domes are too broad and open to permit the formation of phacoliths much thicker than several hundred feet (Fig. 2a). It is obvious that the domes have been eroded so much that the floors of such phacoliths would have been exposed long ago and most

of the igneous body destroyed (Fig. 2b). In no case, however, has even a small part of a floor been observed. Had the main arches of these domes been tightly compressed, there would have been sufficient slippage between beds of the country rock and release of pressure at the tops of the anticlines to produce much thicker bodies (Fig. 2c) as in the case of the Adirondack phacoliths described by Buddington (1929, p. 56). In that event we would expect to find the floor exposed only after considerable erosion (Fig. 2d). (2) If the domes had developed under regional compression and the magma flowed in to form phacoliths, there would have been practically no shearing or slipping between and within beds in these broad, open arches. The writer believes, as will be explained later, that some rock flowage, accomplished by slipping and shearing during the doming, was essential in the formation of the bedding-plane schistosity.

Two principal objections may also be given to the theory that the domes represent upwarps of a single folded sill: (1) The idea of a thick sill or sheet continuous from dome to dome does not seem to be in harmony with the occurrence of zones of intense deformation in the metamorphic rocks between adjacent domes. Such a zone is found between the Croydon and Unity domes where rocks of the highest grade of metamorphism in the area occur. Although the existence of these zones has been recognized for some time (Billings, 1937, p. 532; and Chapman, 1939, p. 155 and 157), their true significance has not been explained. It is quite apparent from field studies that the deformation is a result of regional folding. These zones are so located as to indicate that they are the result of crowding of isolated igneous masses. Regional folding must have disturbed the isolated igneous cores of the domes, and as they were moved relatively to each other they mashed and deformed the weaker metamorphic rocks between them. Were we dealing with a sill or sheet, there could be little relative movement between adjacent domes and certainly not enough to explain the intensity of deformation observed. (2) Were these domes produced by deformation of a sill there would have been little or no shearing parallel to the bedding in the roof rocks. As will be explained later, some shearing and flowage occurred, and this was essential to the formation of the schistosity which parallels the bedding of the metamorphic rocks.

ORIGIN OF FOLIATION IN THE INTRUSIVE ROCKS

Several facts suggest that the foliation of the intrusive cores of the domes is a primary flow structure: (1) the parallel arrangement of flat or elongate inclusions; (2) the parallelism between foliation and contacts; and (3) the marked development of foliation near contacts. It seems

that the effect of doming had much to do with the production of this foliation. As the rock was crystallizing the upward push of the magma below aided orientation.

SCHISTOSE STRUCTURES

The origin of schistose structures is a problem in most metamorphic terranes, and the Claremont-Newport area is no exception. Here two distinct schistosities are recognized in the metamorphosed sedimentary and volcanic rocks. One is parallel to bedding and confined to the rocks immediately surrounding and covering the igneous cores of the Oliverian domes. It will hereafter be referred to as a bedding-plane schistosity. The other is parallel to the axial planes of folds and confined principally to the rocks of the Littleton formation west of the Oliverian domes. This will be referred to as an axial-plane schistosity.

The axial-plane schistosity is believed to have been developed by intense folding, as a result of horizontal compression, during the period of regional metamorphism. In general, its strike is constant over large areas, trending roughly north-northeast, and its dip is steep (60°-90°). For the three following reasons, the axial-plane schistosity must be considered genetically independent of the bedding-plane schistosity.

(1) The axial-plane schistosity is much more uniform in strike and dip than is the bedding-plane schistosity. The marked parallelism between the latter and the intrusive contacts strongly suggests a genetic relationship between the bedding-plane schistosity and the intrusives. Even slight changes in the strike of the bedding-plane schistosity correspond to similar changes in the strike of the igneous contact (Pl. 1).

(2) Bedding and schistosity are essentially horizontal in the vicinity of Unity Mountain, and throughout the rest of the metamorphic rocks east of the Grantham fault bedding and schistosity dip at low angles. The dips are also very low in the remnant of the roof just northwest of Croydon Flat. In all these areas the rocks form portions of the flat or gently dipping dome roofs, and in them only broad, open folds have developed. The gentle flexures observed indicate conclusively the unimportance of regional folding here. It is difficult to conceive of this flat bedding-plane schistosity as forming solely from regional compression where both bedding and schistosity are only slightly warped. The schistosity is also parallel to bedding around the margins of the domes, where the rocks have been more steeply tilted and, as will be considered later, is in part genetically related to the flatter bedding-plane schistosity in the roof rocks.

(3) Crinkling and folding of the bedding-plane schistosity are pronounced in some localities. It has already been shown that large open folds occur in the schistosity and bedding at the north end of the Croy-

don dome. As one proceeds westward across Croydon Mountain, one passes from rocks in which a bedding-plane schistosity is only broadly folded over rocks in which this bedding-plane schistosity is more tightly folded and finally onto rocks (Littleton formation) in which an axial-plane schistosity has formed. It is difficult to determine just where the tightly folded bedding-plane schistosity passes into a true axial-plane schistosity. It appears certain, however, that on the east side of a theoretical transition zone the rocks possess a bedding-plane schistosity with varying degrees of folding. On the west side of this zone, it is equally certain that the rocks possess an axial-plane schistosity. One and a half miles northwest of Grantham Mountain the relationship between bedding and axial-plane cleavage is clear. Here both structures have nearly the same strike, but the angles of dip vary by as much as 50 degrees. A few folds in bedding plunge gently to the south, but the majority plunge 20° NNE., and their axial planes dip steeply to the east. The attitude of the folds here is close to that of the folds in bedding-plane schistosity farther east. One concludes that the crinkling and folding of the bedding-plane schistosity was concurrent with the development of the axial-plane schistosity. The bedding-plane schistosity, therefore, must be distinctly older.

ORIGIN OF BEDDING-PLANE SCHISTOSITY

The schistosity of the metamorphic rocks of the Oliverian domes was considered in an earlier report (Chapman, 1939, p. 155) to have been produced by dynamic and contact effects of the intrusives. In the present study the writer believes that such effects may satisfactorily explain the same structure in the Croydon and Unity domes.

It is believed that for each Oliverian dome an influx of magma, at essentially one horizon, domed the overlying rocks. With the continued raising of these laccolithic arches, the upward force of the magma acted against the weight of many thousands of feet of overlying sediments to compress and thin the roof rocks. This thinning, accompanied by extension or elongation everywhere parallel to the dip of the bedding, caused some shearing and slipping along bedding planes. Shearing served to partially granulate, reorient, and recrystallize many of the minerals and thereby produce an incipient bedding-plane schistosity. Contact metamorphism played a dominant role in the process of recrystallization even after doming. The heat and solutions (water) from the crystallizing cores aided in the prolific development of micaceous minerals, and, controlled by the bedding and incipient schistosity of the rocks, these new minerals oriented themselves to produce a definite schistose structure.

It now remains to be shown that the bedding-plane schistosity is not to be ascribed entirely to thermal and hydrothermal contact meta-

morphism because it has already been stated that some shearing parallel to bedding planes was essential.

According to Harker (1932, p. 35) a foliated structure may be produced purely by contact metamorphism of sedimentary rocks. In such rocks it is supposed that the bedding structure of the sediments controls the growth of new minerals so as to give parallel orientations. Though this process alone may produce a bedding-plane schistosity in thinly bedded rocks like shales, it is doubtful if it could account for schistose and foliated structures in the more massive rocks. The bedding of most of the rocks of the Clough formation (originally sandstones) and Ammonoosuc volcanics (originally tuffs, flows, and breccias) must have been so poor and the rocks so massive that it is doubtful if bedding alone could have exerted much control on the orientation of minerals during contact metamorphism. It is evident, therefore, that shearing was an essential factor in the production of the bedding-plane schistosity. The shearing cannot be attributed to regional compression, at least at the north and south ends of the domes and in the roof remnants where the metamorphic rocks have been scarcely disturbed by folding. It may be attributed, however, to the thinning and stretching of the roof rocks as they were punched upward and domed by the intrusives.

EFFECT OF REGIONAL METAMORPHISM

The Claremont-Newport area was subjected to the conditions of middle-grade metamorphism during the period of folding which postdates the formation of the domes and their bedding-plane schistosity. The intensity of folding varied considerably from place to place. Recrystallization, however, was probably almost universal because it was accomplished in large part by high temperature. Even where folding and shearing were ineffective, the temperature of the rocks was still sufficient for recrystallization.

As has already been shown, folding in and about the domes is not everywhere uniform. The intensity is much greater on the west flanks than at either the north or south ends. Such is the case for nearly all the Oliverian domes. In general, the rocks on the western flanks of the Croydon and Unity domes are tightly folded (almost isoclinal), but at the north and south ends gentle folding is the rule. Furthermore, it is significant that the intensity of folding increases notably to the west at the north end of the Croydon dome. The same relationship holds for both the north and south ends of the Unity dome but is not so strikingly demonstrated in the field. It is concluded, therefore, that the plutonic cores of the domes acted as buttresses and locally protected the metamorphic rocks from intense deformation.

An analogy may be made between the effect of these resistant cores and the development of certain eyed structures in metamorphic rocks where schist layers wrap partly around large crystals of garnet or magnetite. The large crystal has either forced apart the folia or caused the folia to partly wrap around it. There results a lens-shaped opening between the folia with the resistant crystal at its center. In this manner the strong crystal apparently relieves the pressure at the corners of the eye and permits these spaces to be filled with some more soluble mineral such as quartz.

In the case of the Croydon and Unity domes, it is believed that orogenic pressure caused many of the rocks to yield by folding. Tightly closed folds, with essentially horizontal axes and steep axial planes, formed along the western flanks. This caused the bedding-plane schistosity, which probably originally dipped gently west, to be tightly folded and to assume a steep dip. In places where bedding is poor or absent it is impossible to determine whether one is dealing with the tightly folded, bedding-plane schistosity or with a true axial-plane schistosity. It seems very likely, however, that where the bedding plane schistosity (originally gently inclined) was tightly folded, as along the west flanks of the domes, the rocks were again considerably recrystallized. The folded bedding-plane schistosity must have so controlled the orientation of minerals formed in this second period of recrystallization that the final schistose structure follows the earlier bedding-plane schistosity.

The marked decrease in intensity of folding at the north and south ends of the domes seems to be in accord with the belief that the igneous cores resisted, to a great extent, the orogenic forces directed against their flanks. The rocks at the ends of the domes were not permitted to be so tightly squeezed as were those along the flanks. Just as the large crystal in the metamorphic rock relieved the pressure in the corners of the eyed structure, so the plutonic cores may have relieved the pressure at the ends of the domes.

Probably most of the Littleton formation west of the domes was so far removed from the Oliverian intrusives that it never developed a bedding-plane schistosity characteristic of those rocks nearer the intrusives. Although schistosity in the Littleton formation is very locally parallel to bedding where observed, no evidence of an extensive bedding-plane schistosity was noted. Field studies indicate that regional compression folded these rocks and metamorphosed them for the first time, developing a schistose structure of the axial-plane type.

The intense folding and crumpling between the Croydon and Unity domes is attributed to crowding of the two isolated igneous cores during regional folding. It is believed that here also the bedding was tightly

folded, the rocks recrystallized, and the resulting schistose structure developed parallel to the bedding. If all these rocks possessed a bedding-plane schistosity before folding, this structure probably controlled the orientation of minerals during regional metamorphism. So intense were the conditions of temperature and pressure here that rocks of the high-grade zone of metamorphism (sillimanite schist) formed. The present metamorphic character of the rocks between the domes as well as most of the Littleton formation west of the domes is perhaps to be attributed principally to regional metamorphism. The diversity in orientation of nonrotated staurolite porphyroblasts in the Littleton formation indicates that this mineral crystallized late during regional metamorphism.

MT. CLOUGH PLUTON

In the extreme eastern part of the area is the Mt. Clough pluton, a highly elongate intrusive body composed of Bethlehem gneiss. Although a study of the gneiss in the present area contributes little to our knowledge of the pluton, it seems advisable to give a brief discussion of the body here. The Mt. Clough pluton extends in a north-south direction for at least 85 miles, and its east-west dimension is known to range from half a mile to 7 miles. The gneiss has a prominent foliation which is, in general, best developed near the walls of the intrusive, and this foliation throughout the entire length of the body is remarkably concordant with the boundaries of the pluton. Most turns in the contact are reflected by corresponding turns in the strike of the foliation. Because of the Grantham fault this marked parallelism between contact and foliation is much less striking in the present area than it is in the Mascoma quadrangle to the north or the Bellows Falls quadrangle (F. C. Kruger, personal communication) to the south.

The pluton, as has already been shown (Billings, 1937, p. 536; Chapman, 1939, p. 161), is a huge sill-like body of late Devonian age which, throughout most of its extent, was intruded at the base of the Littleton formation. Everywhere north of the Sunapee quadrangle it separates high-grade schist of the Littleton formation on the east from metamorphosed Ammonoosuc volcanics on the west. Locally, on the west side of the pluton, parts of the intervening Clough and Fitch formations are preserved in contact with the gneiss. It is believed that these formations and perhaps large slices of the Ammonoosuc volcanics and Littleton formation were pushed ahead of, or engulfed by, the intrusion.

The most pronounced cross-cutting observed anywhere along the western boundary of the pluton is here in the Claremont-Newport area. Northwest of Newport, where the base of the Bethlehem gneiss diverges from the Grantham fault (Pl. 1), the Clough formation is in contact with

the gneiss. Just south of Mt. Tug the gneiss is in contact with the upper part of the Ammonoosuc volcanics. Between this point and the intrusive contact near Unity (Pl. 1), the gneiss must cut across the entire Clough formation and many hundred feet of the Littleton formation. Thus, from Unity to the southern boundary of the area and even in the Bellows Falls quadrangle (Kruger and Linehan, 1941, p. 642), the base of the intrusive lies in the Littleton formation, many hundred feet above the top of the Clough formation.

This cross-cutting relationship is believed to be due to intrusion and not to later faulting. The contact between the gneiss and metamorphic rocks is best observed nearly due west of Newport village, on Mt. Tug (Pl. 1). Here the Bethlehem gneiss rests directly on the Clough formation, and the contact strikes N. 20° E. and dips 25° SE. The foliation of the gneiss as well as the schistosity of the underlying Clough formation is, within a few degrees, parallel to the contact. Linear features near the contact are due principally to crinkling and plunge about 24° SE. This direction corresponds closely to that for similar linear features in much of the surrounding area. In general the Bethlehem gneiss is a little finer-grained within a few yards of the contact. This is not only true on Mt. Tug but also for several miles farther south. This fact may not be of any great significance, however, as the grain size of the body has been observed to vary considerably, even near the center of the pluton. No features characteristic of faulting were observed along the contact, nor is the evidence indisputably in favor of an intrusive contact here. Nevertheless, it seems more logical to consider the contact as intrusive rather than a recrystallized fault.

Billings (1937, p. 536) has shown that in the northern part of the pluton the walls are probably vertical. Farther south, however, the average dip is 45° E. (Fowler-Lunn and Kingsley, 1937, p. 1376; Chapman, 1939, p. 163), and in the southern part of the Mascoma quadrangle the intrusive sheet must have originally wrapped up over the north and east sides of the Croydon dome. Displacement along the Grantham fault brought down the Bethlehem gneiss, which covers the eastern part of the dome, and preserved it from erosion. This interpretation is in part substantiated by the nearly horizontal foliation of the gneiss which extends from a point north of Grantham nearly to Croydon Flat. Here the intrusive sheet is nearly horizontal as it rests on top of the relatively downfaulted portion of the Croydon dome.

SUMMARY

A study of the Croydon and Unity domes throws further light on the origin of the Oliverian domes. Evidence favors interpreting the domes

as laccolithic bodies which were intruded into flat-lying Paleozoic rocks probably in late Devonian time. Primary flow layers formed in the igneous rocks parallel to intrusive contacts during intrusion, and a bedding-plane schistosity developed contemporaneously in the surrounding Paleozoic rocks. During the period of orogeny which followed, an axial-plane schistosity was superimposed on the rocks west of the domes. The igneous cores of the domes acted as buttresses and locally protected the immediately surrounding rocks from the deformation. Folding of the dome rocks was most intense on the west flanks and least intense at the north and south ends and in parts of the roofs. The intensity of folding in many places increases with distance from the igneous core.

Further study of the Mt. Clough pluton shows that it apparently continues to the south essentially as a concordant sheet. It is known to cross-cut the bedding locally and to rise from the base of the Clough formation to several hundred feet up into the Littleton. However, this cross-cutting is not great if one considers it in relation to the total length of the body.

REFERENCES CITED

Billings, M. P. (1937) *Regional metamorphism of the Littleton-Moosilauke area, New Hampshire,* Geol. Soc. Am., Bull., vol. 48, p. 463-566.

Buddington, A. F. (1929) *Granite phacoliths and their contact zones in the northwest Adirondacks,* N. Y. State Mus., Bull. 281, p. 51-107.

Chapman, C. A. (1939) *Geology of the Mascoma quadrangle, New Hampshire,* Geol. Soc. Am., Bull., vol. 50, p. 127-180.

——— (1941) *The tectonic significance of some pegmatites in New Hampshire,* Jour. Geol., vol. 49, p. 370-381.

Fowler-Lunn, Katharine, and Kingsley, Louise (1937) *Geology of the Cardigan quadrangle, New Hampshire,* Geol. Soc. Am., Bull., vol. 48, p. 1363-1386.

Hadley, J. B. (1942) *Stratigraphy, structure, and petrology of the Mt. Cube area, New Hampshire,* Geol. Soc. Am., Bull., vol. 53, p. 113-176.

Harker, Alfred (1932) *Metamorphism,* Methuen and Co. Ltd., London.

Holmes, Arthur (1928) *The nomenclature of petrology,* 2d ed., Thomas Murby and Co., London.

Kruger, F. C., and Linehan, S. J. (1941) *Seismic studies of floored intrusives in western New Hampshire,* Geol. Soc. Am., Bull., vol. 52, p. 633-648.

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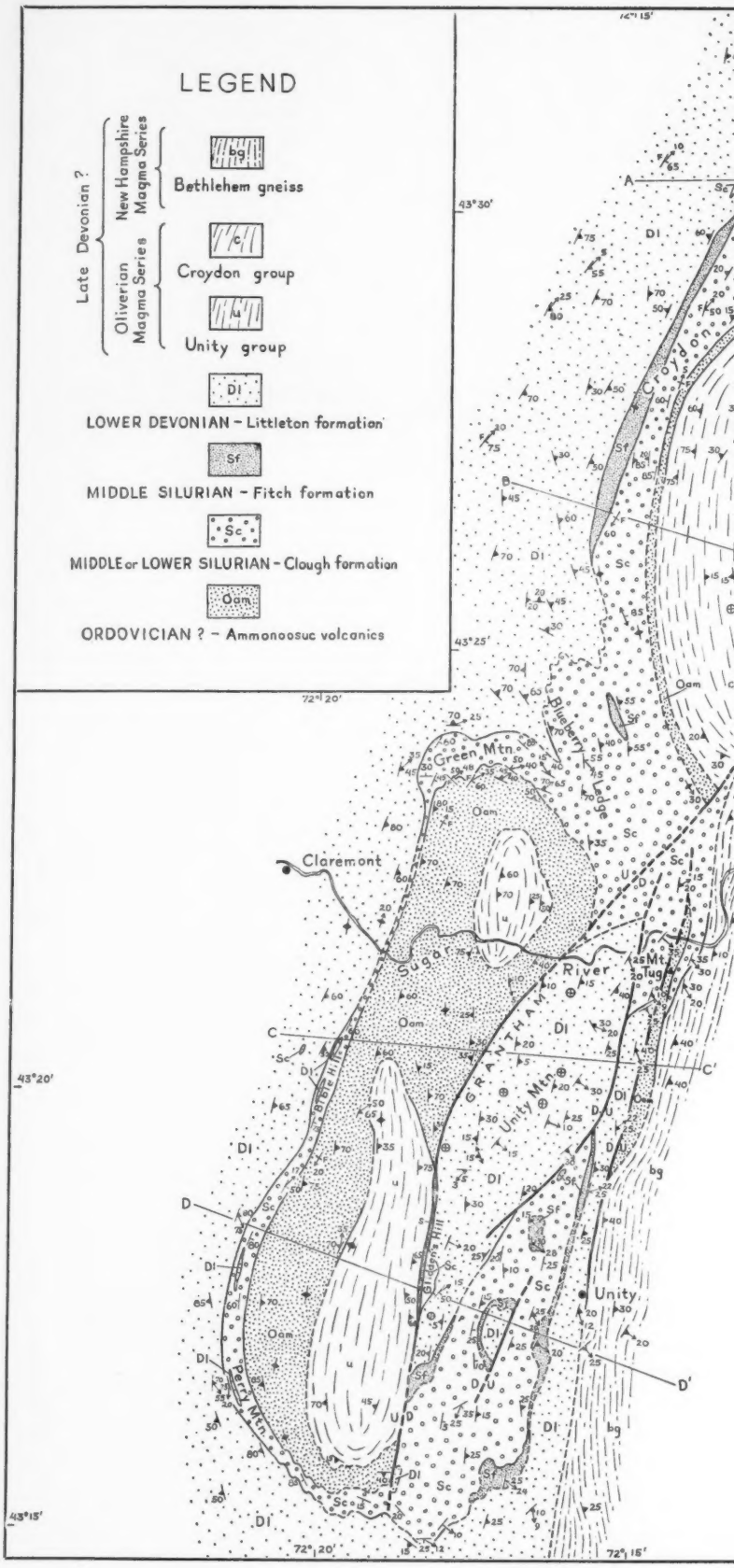

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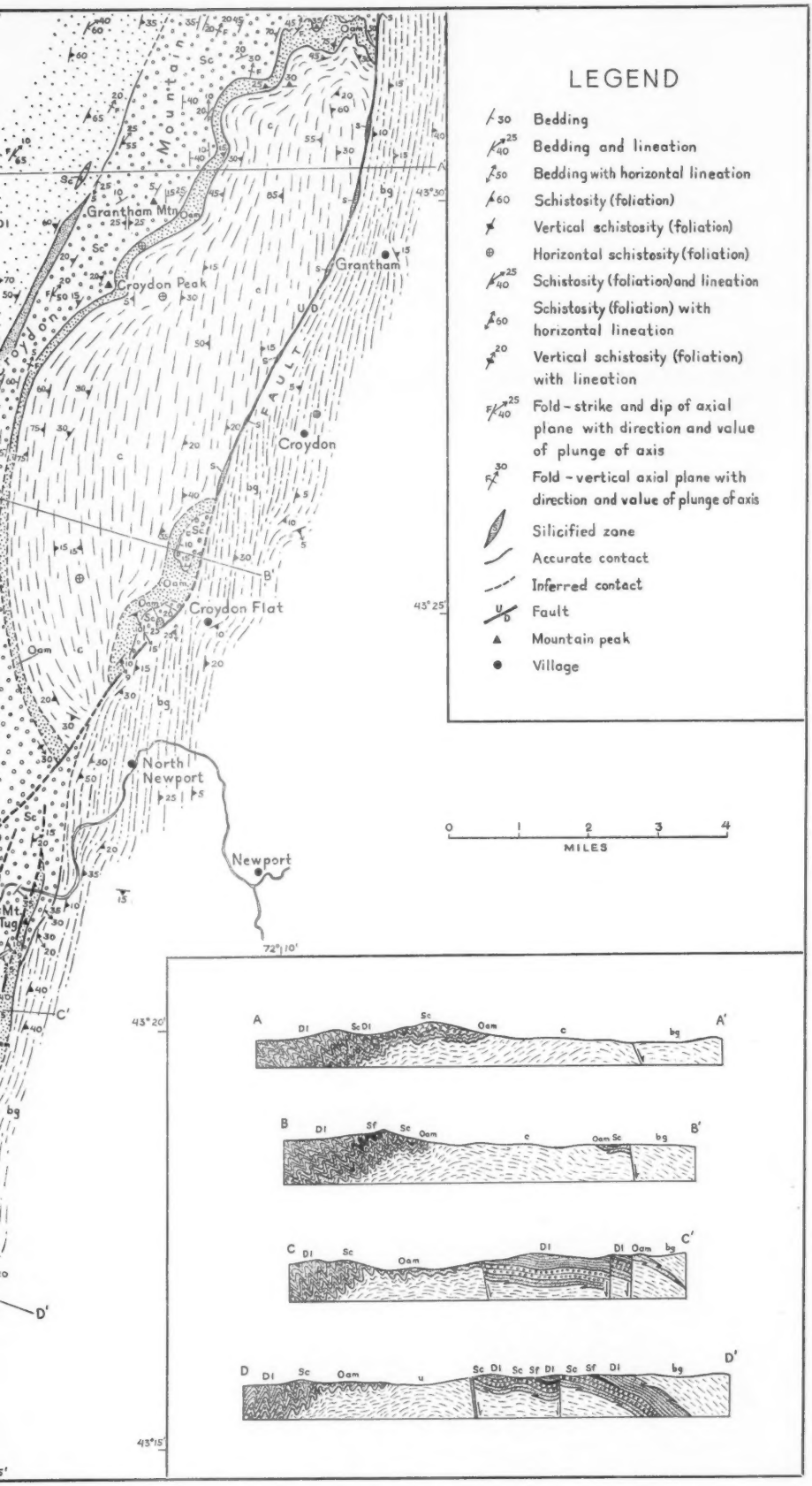

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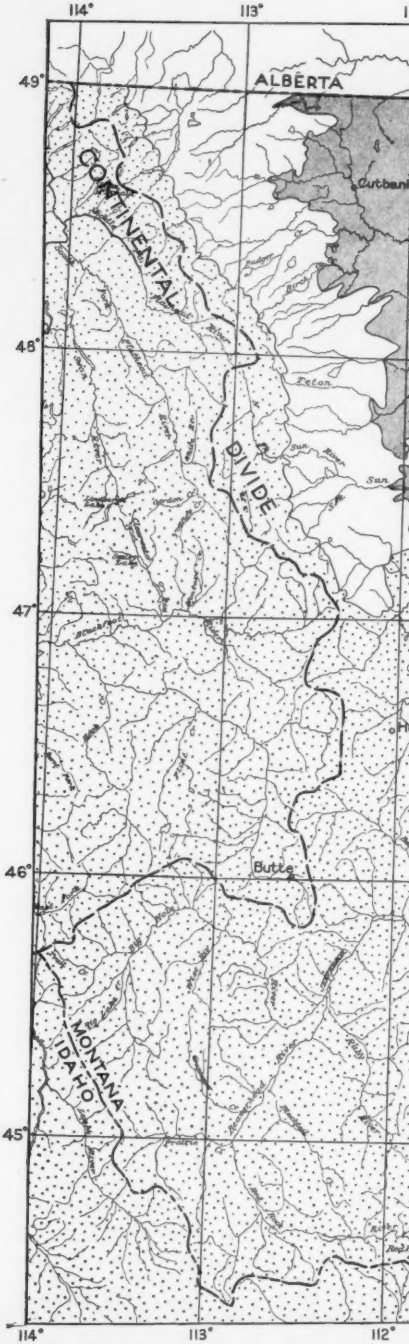

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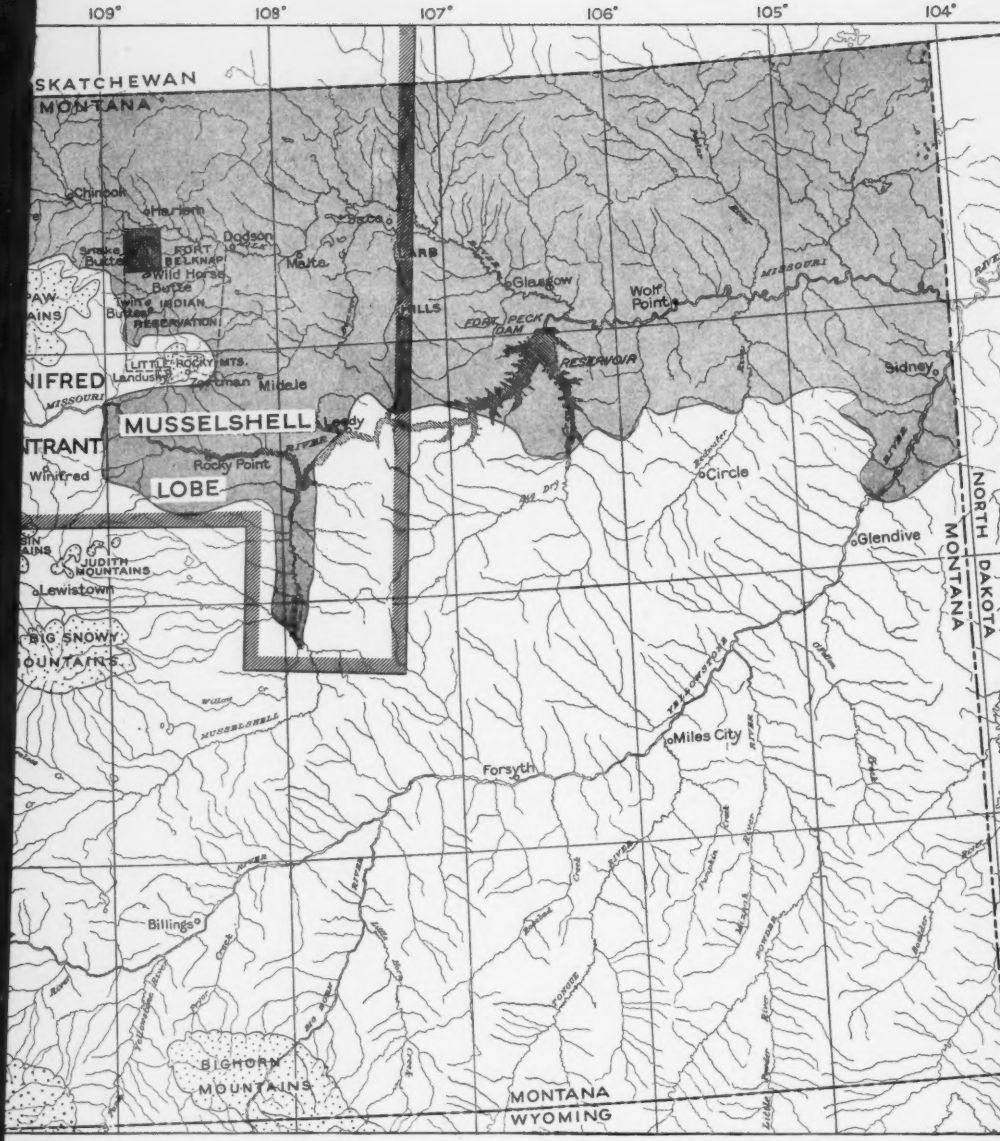
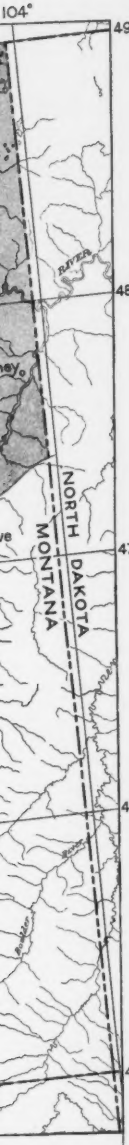


AINES AND GLACIOLACUSTRINE DEPOSITS OF THE KEEWATIN GLACIER IN MONTANA



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SNAKE BUTTE BOULDER TRAIN AND RELATED GLACIAL PHENOMENA, NORTH-CENTRAL MONTANA

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ABSTRACT

Glacial striae on shonkinite exposures at Snake Butte, a prominent intrusive in the plains south of Harlem, Montana, trend southeastward. A train of shonkinite boulders, many of them enormous and some of them striated and polished, extends southeastward 50 miles from Snake Butte across the plains of the Fort Belknap Indian Reservation. The indicated southeastward movement of the ice is explained as a diversion of the southward-moving Keewatin glacier as it piled against the north side of the Bearpaw Mountains, developing a downward gradient toward the southeast.

INTRODUCTION

Large blocks of igneous rock that were transported during the glacial epoch southeastward from the prominent sill cropping out at Snake Butte, 10 miles south of Harlem, Montana, (Pls. 1, 2, 3; Figs. 1, 4) were noted by Alden (1932, p. 107) in his discussion of glacial phenomena in the plains adjacent to the Little Rocky Mountains. During the writer's more recent geologic studies of the Fort Belknap Indian Reservation, the Little Rocky Mountains, and neighboring areas, the distribution of these blocks was mapped and a number of striated rock exposures on Snake Butte were examined. The blocks form a well-defined boulder train extending in the same southeasterly direction as the striae and, in one stage, at least, the Keewatin glacial ice therefore moved southeastward across the plains north of the Little Rocky Mountains. It has been assumed heretofore that the direction of motion was southwesterly. The significance of the new data in relation to available information on the topography and glacial geology of the Montana plains has been considered and is here discussed.

ACKNOWLEDGMENTS

The published findings of Calhoun (1906), Alden (1932), and other workers have been utilized, and the writer is indebted to several of his associates for information and valuable suggestions received during the preparation of this paper. The manuscript has been read critically by W. C. Alden, G. F. Loughlin, H. D. Miser, and W. H. Bradley.

SNAKE BUTTE

TOPOGRAPHIC FORM

Snake Butte (Pl. 2; Fig. 1) occupies a roughly elliptical area of about 1½ square miles in Ts. 30 and 31 N., R. 22 E., 10 miles south of Harlem. Its longest dimension, about 2½ miles, extends northeastward. The butte is highest in its northeast part where, with an altitude of 3190 feet above sea level, it stands about 400 feet above the surrounding plains. The top of the butte slopes gently southwestward and a knob about 100 feet high stands at its southwest end. A long, high cliff extends along the southeast side, facing the plains.

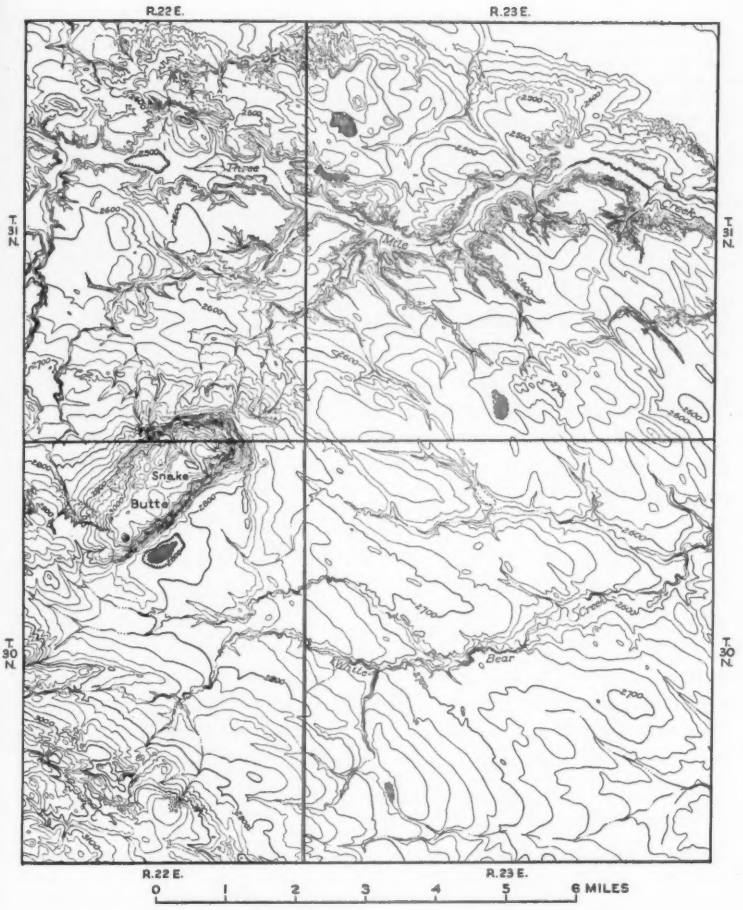

FIGURE 1.—*Map of Snake Butte and adjacent plains*

Showing dominant southeastward trends in topography. Contours (interval 20 feet) by A. F. Dunnington, of the General Land Office, United States Department of the Interior. Cross-ruled areas are lakes.

IGNEOUS ROCK

Most of the igneous rock exposed at Snake Butte is dark, medium-grained shonkinite, containing approximately equal amounts of feldspar and dark minerals. A mafic syenite is exposed in a small topographic depression near the center of the butte. Cathcart (1922) furnishes the following petrographic description of the shonkinite:

"Orthoclase is the predominant feldspar in the rock and augite the most abundant dark mineral. Some albite is graphically intergrown with the orthoclase and some of the augite is bordered by aegerite. Biotite is plentiful and olivine is prominent in some specimens. Apatite and magnetite are present as accessory minerals; sericite, kaolin, and a zeolite (probably natrolite) are of secondary development."

The shonkinite forms a sill, limited to Snake Butte; on all sides it is in contact with Cretaceous sedimentary strata into which it was intruded in Tertiary time. In the northeastern part of the intrusive columnar jointing is well developed. Columns 10 feet and more in diameter, nearly vertical and cut by horizontal joints (Pl. 2, fig. 2), are arrayed impressively in the cliff along the southeast side, at the base of which are great accumulations of talus. Where exposed on the top of the butte, the tops of the columns, most of which are five-sided, form a mosaic covering several acres. A few large groups of leaning columns which have dropped down immediately in front of the cliff and a long, deep northeastward-trending fissure in the top of the butte not far back of the cliff probably signify either glacial action or landsliding.

GLACIAL STRIAE

At several places on the top of Snake Butte the exposed shonkinite bedrock is distinctly striated and a number of compass readings indicate that the directions of striation range from S. 37° E. to S. 52° E., averaging S. 45° E. The striae shown on Figure 3 of Plate 3 trend S. 37° E. and are about a quarter of a mile N. 15° E. of the prominent knob on the butte near its southwest end. Most of the striae observed, however, are on the northeastern part of the butte, where they are best preserved on rock surfaces that are protected from weathering under several large boulders.

Owing to the softness of most of the bedrock in the part of the Great Plains occupied by drift from the Keewatin ice, glacial striae are rare, having been reported at only three other localities in north-central Montana (Fig. 4). Stebinger (Alden, 1932, p. 105, Pl. 39) observed striae bearing southward on Signal Butte, north of Havre, Montana, on a ledge reported to be somewhat dislocated. Alden (1932, p. 101) describes striae trending S. 4° W. and S. 8° E. on top of Boxelder Butte, on the northwest side of the Bearpaw Mountains. Pecora (personal communication, 1940) says that striae on the southwest slope of Black Butte, sec. 10, T. 32 N., R. 15 E., near Pacific Junction, 4 miles west of Havre, bear S. 30° E. These markings are on a boulder in the volcanic breccia that forms the bedrock at this locality.

FIGURE 1. THE BUTTE AS SEEN FROM EAST

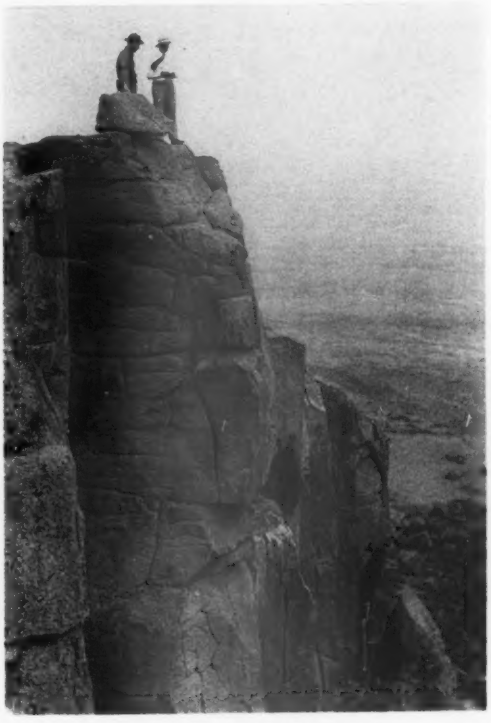

FIGURE 2. COLUMNS OF SHONKINITE
(Photograph by H. D. Miser)

SNAKE BUTTE

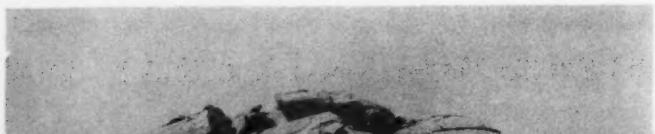

FIGURE 1. SHONKINITE BOULDER NEAR MALTA-ZORTMAN ROAD

FIGURE 2. SNAKE BUTTE BOULDER TRAIN NEAR HARLEM-HAYS ROAD

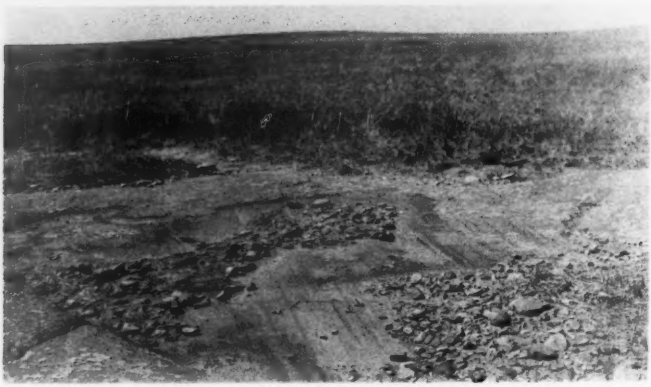

FIGURE 3. TOP OF SNAKE BUTTE
Showing glacial striae and cobbles

GLACIAL BOULDERS, COBBLES, AND STRIAE

GLACIAL DRIFT

AREAL EXTENT, UNDERLYING BEDROCK, DRIFTLESS AREAS

The drift deposited by the Keewatin ice sheet occupies about 38,000 square miles in the Montana plains (Pl. 1), a region where most of the exposed bedrock is shale and sandstone of Mesozoic and Tertiary age. The southern margin of the drift extends in a general westerly direction from North Dakota to a point on the Missouri River near Great Falls, where its general course turns northwestward. In this distance of more than 300 miles the most prominent departures from the general westerly course are the lobe in the vicinity of the Musselshell River and the large re-entrant north of Winifred in which are included two isolated mountain groups—the Bearpaw and Little Rocky Mountains. The Great Falls lobe, described by Calhoun and Alden, lies between the Winifred re-entrant and the belt of high plains adjacent to the Rocky Mountain front. Calhoun's map (1906, Pl. 1) shows three isolated driftless areas at the Sweetgrass Hills.

RELATION TO RELIEF

The part of Montana occupied by drift from the Keewatin ice has low relief with a general easterly slope from the base of the Rocky Mountains to the North Dakota boundary. The belts of low country are followed by the Missouri River and its larger tributaries, the Teton, Marias, Milk, Judith, Musselshell, and Yellowstone rivers; the higher tracts lie between these principal streams, adjoin the Rocky Mountain front, and surround the isolated mountainous areas. The Sweetgrass Hills are completely surrounded by glacial drift and the Bearpaw and Little Rocky Mountains are nearly so, but all other mountainous parts of Montana lie outside the area occupied by drift from the Keewatin ice.

The morainal material deposited by the Keewatin ice sheet apparently reaches an altitude not greatly exceeding 4000 feet above sea level in the neighborhood of the Bearpaw and Little Rocky Mountains. According to Reeves (1924, p. 12), however, the southern limit of the drift as mapped in the vicinity of the Bearpaw Mountains does not represent the limit of the ice advance but the limit of areas where the bedrock is wholly concealed beneath drift. Pecora (personal communication, 1940) observed two erratic boulders farther south at altitudes above 5000 feet in the Bearpaw Mountains. One of these is a block of pre-Cambrian garnet-bearing gneiss on a ridge in the southern part of sec. 8, T. 28 N., R. 16 E., east of Boxelder Creek, near its head. The other is a block of granite pegmatite on the divide near the head of Godfrey Creek, about 500 feet higher than a spring about half a mile southwest of the boulder. Glacial erratics

were noted by Alden (1932, p. 80, 81, 99, 100) at altitudes as high as 3900 feet above sea level on the north side of the Bearpaw Mountains and as high as 3850 feet on Twin Buttes, northwest of the Little Rocky Mountains. On the north side of the Little Rocky Mountains drift was found by him 3500 feet above sea level near Lodgepole and higher than 4000 feet on outlying buttes southeast of these mountains. On the south side of the Little Rocky Mountains, near Landusky, the altitude of the drift margin is approximately 3400 feet. The interior of the Little Rocky Mountains shows no evidence of glaciation.

AGE

Areas covered by older (Iowan or Illinoian?) and younger (Wisconsin) deposits of drift have been mapped by Calhoun (1906, Pl. 1) and Alden (1932, Pl. 1) in northern Montana. Location of the boundary between these areas east of the Fort Belknap Indian Reservation, however, is stated by Alden (1932, p. 79) to be "largely hypothetical." No separation of older and younger drift seems to be possible in the vicinity of the Reservation and only the largest area covered by the ice sheet is considered in most of the present discussion. The possibility of earlier or later, less extensive invasions does not appear to affect the interpretations presented.

CHARACTER AND SOURCES

General statement.—The glacial drift deposited on the Montana plains by the Keewatin ice is partly composed of rock that has been transported long distances. Foreign material was brought by the glacier from two broad regions: (1) the Canadian Shield region of pre-Cambrian and Paleozoic rock exposures, and (2) the Great Plains. Each region contributed characteristic rock types that, as a rule, are easily recognized in the boulders, cobbles, and pebbles of the drift.

Materials from the Canadian Shield.—The drift contains erratics of granite and other crystalline rocks, including metamorphic rocks, like the types occurring in the pre-Cambrian of the Canadian Shield. Some limestone erratics were derived from the nearly continuous belt of exposed Paleozoic rocks that extends northwestward along the margin of the Canadian Shield from southern Manitoba to the District of Mackenzie. Corals obtained from a limestone boulder in the drift along the main county road about 16 miles north of Harlem were examined by G. A. Cooper and Edwin Kirk (oral communication, 1940), who reported them to be "probably middle Silurian." The crystalline and other rock materials derived from the Canadian Shield are stated by Tyrrell (1898), Hopkins (1923), Rutherford (1928), and

others to have moved in southerly directions across the Provinces of Alberta, Saskatchewan, and Manitoba. Apparently the drift of north-central Montana is derived from north-northeasterly sources.

Materials from the Great Plains.—The drift contains considerable gravel (Pl. 3, fig. 3) picked up by the ice in passing over the extensive Swift Current, Cypress Hills, and Wood Mountain Tertiary gravel deposits of southern Saskatchewan, described by McConnell (1885) and by Fraser and others (1935, p. 54-58), and over the Flaxville gravel of northern Montana, described by Collier and Thom (1918). The gravels of these deposits and the drift materials derived from them consist chiefly of smoothly rounded pebbles and cobbles, commonly of maroon and buff quartzite, siliceous argillite, and igneous rocks; in Tertiary time streams transported these sediments onto the Great Plains from the pre-Cambrian Belt formations of the Rocky Mountains.

The Canadian Shield materials and Tertiary gravels are intermixed in the coarser drift with a subordinate amount of sandstone and other rocks that crop out in the Great Plains. The predominance in the coarse drift of rocks of Canadian Shield and Tertiary gravel types is due to their durability, which is greater than that of most of the other rocks. Probably the softer, less resistant rocks were largely converted to "glacial flour" before they had traveled far, furnishing much of the finely divided material that forms a large proportion of the till in the Montana plains.

Snake Butte boulder train.—Almost all of the rock that forms the boulder train extending at least 49 miles southeastward from Snake Butte is shonkinite identical with that of the intrusive, but a few blocks are mafic syenite like that exposed near the center of the butte. No similar rocks occur in the drift elsewhere in the surrounding plains and the only other known outcrops of shonkinite within a radius of 100 miles of Snake Butte are those of the Bearpaw Mountains, reported by Pecora (1941), and those of the Highwood Mountains, described by Weed and Pirsson (1895) and by Hurlbut (1939). The Snake Butte intrusive is therefore the only possible source of the boulders.

A straight line drawn approximately S. 49° E. from the center of Snake Butte closely follows the course of the boulder train, notwithstanding the slight bend (Fig. 4) in its northwestern part.

The Snake Butte boulder train rests on a nearly horizontal, flat surface. The boulder train is somewhat irregular in width with a maximum of 2½ miles. Its greatest width is thus approximately equal to the length of the shonkinite exposure at Snake Butte. This length

is normal to the axis of the boulder train and probably, therefore, controlled its width.

The nearly straight and narrow form of the Snake Butte boulder train differs from that of some boulder trains in Wisconsin (Buell, 1895), New England (Shaler, 1895, p. 185; Taylor, 1910, p. 747-752; Antevs, 1922, Pl. 6; Woodworth, et al., 1935, Pl. 6), and the Baltic region (Hausen, 1912; Sauramo, 1929), which broaden out as fanlike zones away from the parent ledges.

Most of the shonkinite boulders are 10 feet or less in diameter, but some of them are enormous. The plat of T. 26 N., R. 27 E., prepared in 1894 by Charles H. Redfield of the General Land Office, indicates one measuring 20 by 50 by 75 feet. This boulder (Pl. 3, fig. 1), which resembles a haystack when seen at a distance, is broken into blocks that are clearly the segments of several adjacent columns of shonkinite. A similar boulder, about half a mile to the east, is slightly smaller and another, half a mile north, is somewhat larger. These three boulders, believed to be the largest in the train, are about 36 miles from Snake Butte. United States Coast and Geodetic Survey B. M. J-66 (altitude 2940.768) on the east side of the Malta-Zortman road, is set in a 12-foot shonkinite boulder. Unlike the Iron Hill boulder train of Rhode Island as described by Shaler (1893, p. 198), the Snake Butte boulder train shows no progressive decrease in the average size of the blocks with distance from their source, nor is there any progressive increase in the average distance between the blocks. The most southeasterly boulder that has been found lies 49 miles from Snake Butte in the NE. ¼ sec. 24, T. 25 N., R. 28 E., 2½ miles east of Midale cemetery. This block, in contrast to all others observed, was discovered only after a careful search of the neighborhood. The blocks of shonkinite were quickly found at all places in the boulder train between Snake Butte and a point about 5 miles northwest of this boulder; but none have been found beyond it to the southeast nor in the area to the south.

The blocks of shonkinite are scattered in some profusion along most of the roads and trails crossing the boulder train (Pls. 2, 3). At many places along it, as at the Harlem-Hays road, (Pl. 3, fig. 2) they form a large proportion of the glacial boulders scattered over the surface. At other places only a few are present, but even at such localities their characteristic dark-gray color and blocky shapes render them conspicuous. They show little weathering and, though commonly they are well striated and polished, they have been subjected to less wear than the boulders of foreign material in the drift. Occasional blocks many miles from Snake Butte have retained the five-sided shape of the columns at the northeast end of the butte.

ROCHES MOUTONNÉES? AND ROCK BASINS?

The southeast trend of most of the minor streams and the elongation in the same direction of many low hills are striking features of the plains crossed by the Snake Butte boulder train (Fig. 1). As some of these hills, at least, are composed of Cretaceous shale and sandstone, thinly covered with boulder-bearing drift, it is possible that they are roches moutonnées, though this has not been established. Several small basins containing ponds or shown by depression contours in Figure 1 are also elongated in southeasterly directions and may be rock basins caused by glacial erosion.

PARALLEL TRENDS OF GLACIAL, TOPOGRAPHIC, AND BEDROCK FEATURES

The trend of most minor streams and the prevailing elongation of hills and basins in the Fort Belknap plains are approximately parallel to the general course of the adjacent part of the drift margin near the Little Rocky and Bearpaw Mountains. They are likewise nearly parallel to the trend of a morainal belt mapped by Alden (1932, Pl. 1) in T. 31 N., R. 23 E., about 7 miles northeast of Snake Butte (Fig. 4), and to the Snake Butte boulder train and striae. As the margin of the drift and the position of the morainal belt may represent, respectively, the margin of the ice sheet at its maximum stage and at a later recessional stage, Alden (1932, p. 107) suggested that the shonkinite boulders might have been carried southeastward by ice floating on water coursing along the glacial margin at one stage in its retreat northeastward. At the time this suggestion was made the southeast-trending striae on Snake Butte were not known and the distribution of the shonkinite boulders had not been mapped. It now appears certain that the boulder train and striae indicate a local direction of movement in the last ice sheet to enter this part of Montana. The southeastward trends in the topography may therefore have resulted partly from the erosive action of the ice. However, Plate 1 and many detailed maps that have been consulted show that a southeast trend of minor streams is characteristic of much of the eastern half of Montana, in unglaciated as well as in glaciated areas. It is therefore possible that the topographic trends shown on Figure 1 are due to some regional factor, or combination of factors, not primarily related to glaciation.

Possibly the bedrock structure of the Montana plains was a factor, for the trend of many of the structural axes mapped by Dobbin and Erdmann (1934) is southeastward; in addition, the writer's investigation indicates that all but one of several known faults in the Cretaceous shale and sandstone in the northern part of the Fort Belknap Indian Reservation trend southeast.

THE ICE SHEET

REGIONAL ADVANCE OF ICE MARGIN

That the Keewatin ice sheet advanced into Montana from the north-northeast is inferred from the derivation of some drift constituents from the Canadian Shield. In harmony with this are the southerly direction

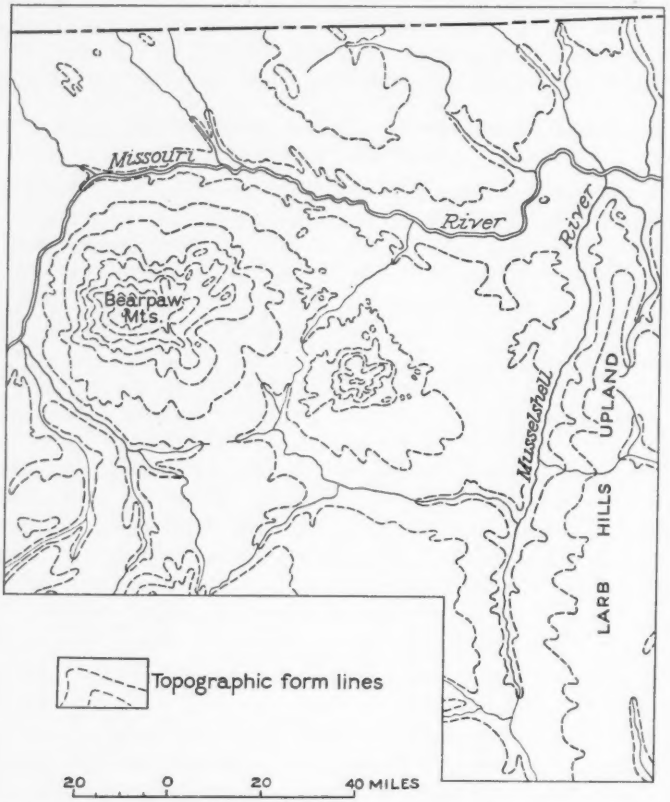

FIGURE 2.—*Map of Bearpaw Mountains, Little Rocky Mountains, and surrounding territory, showing probable preglacial configuration*

of striation at Signal Butte (Fig. 4) and the direction of glacial motion suggested by the prolongation of the driftless areas mapped by Calhoun (1906, Pl. 1) on the south-southwest sides of each of the three former nunataks at the Sweetgrass Hills (Pl. 1). These driftless areas may repre-

sent glacial eddies in which the glacier at the beginning of the ice invasion became stagnant, or nearly so, on the south-southwest side of each nunatak and remained relatively inactive on their leeward sides until the glacier disappeared. If the earliest ice did not carry much rock debris and if little or no debris was added later in those positions, the amount of morainal material that accumulated there might be so small as to have escaped observation. For the same reason it is possible that the area included in the Winifred re-entrant (Pl. 1) was also partly covered by ice.

DIRECTIONS OF GLACIAL MOTION

The phenomena associated with the maximum advance of an ice sheet are not entirely separable from those originating from later oscillations or during retreat; moreover, the exact directions of ice movement when the glacier existed in north-central Montana are known at only a few places. In other regions, as in Wisconsin, there are many evidences of local variations in direction of movement, such as crossed striae. Consequently only a general impression of the deployment of the ice as the glacier approached its farthest southern limit in the vicinity of the Bearpaw and Little Rocky Mountains can be formed from the available data. The most significant information at hand is yielded by the Snake Butte boulder train and the associated striae on Snake Butte.

If more than one ice invasion occurred in this region, as is suggested by Calhoun and Alden, the Snake Butte phenomena probably are due to the last invasion. Nevertheless, in proving that in one stage, at least, the glacier moved southeastward across the Fort Belknap Indian Reservation, these phenomena offer a basis for belief that motion in the same direction would occur whenever the ice entered northern Montana from the direction of the Keewatin center and advanced to the vicinity of the Bearpaw and Little Rocky Mountains. The southeastward direction indicated is that which would probably be taken whenever the ice sheet, spreading toward the south-southwest, was obstructed and turned aside by the high ground in and adjacent to these mountains.

The southerly glacial movement indicated by the striae at Signal Butte (Fig. 4) and by the prolongation of the driftless areas mapped by Calhoun (1906, Pl. 1) southwest of each of the three former nunataks at the Sweetgrass Hills (Pl. 1) suggests that the Great Falls lobe, described by Calhoun (1906) and Alden (1932, p. 101), resulted from a southwesterly movement across the plains between the Bearpaw Mountains and the high land adjacent to the Rocky Mountain front. The southeastward movement indicated by the striae at Black Butte and by the phenomena associated with Snake Butte suggests that the high ground in the vicinity of the Bearpaw Mountains deflected a part of the glacier and caused this

part to move down the preglacial Missouri River valley, now occupied by Milk River. Such a split in the glacier is indicated by the Winifred re-entrant in the margin of the drift.

The presence of a few erratics at altitudes above 5000 feet in the Bear-

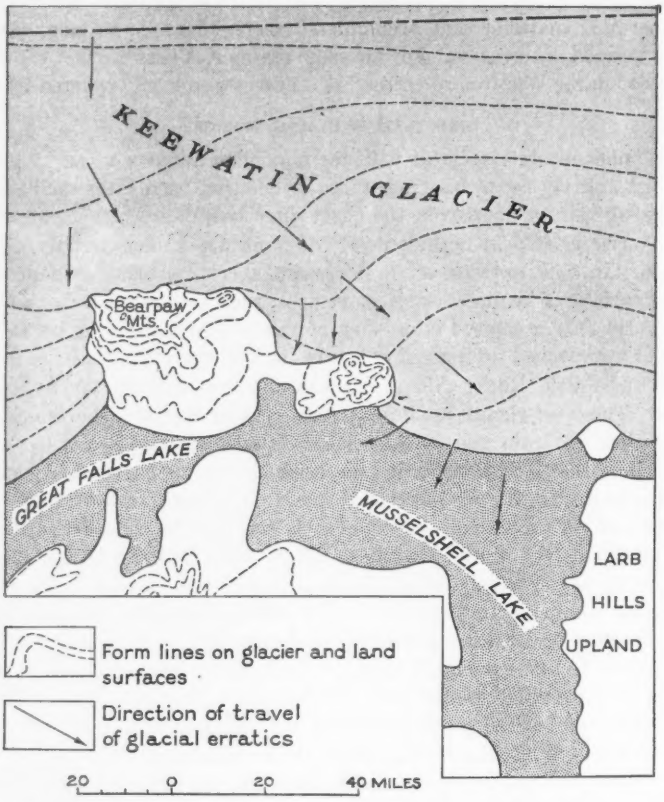

FIGURE 3.—*Same area as Figure 2, showing interpretation of glacial phenomena*

paw Mountains may be due to piling of the upper part of the ice sheet, which probably carried less rock debris than the lower part, against the north side of the Bearpaw Mountains. A thick accumulation of ice north of the Bearpaws may have caused a southeastward gradient of the ice surface northeast of the Bearpaws and Little Rockies, somewhat in the manner shown on Figure 3; this gradient may, for a time, have been suffi-

ciently steep to account for the southeastward glacial movement indicated by the Snake Butte boulder train and striae. The results of studies by Shaler (1893), Hess (1933, p. 114) and other writers indicate that the basal part of a deep glacier is highly mobile, except perhaps where

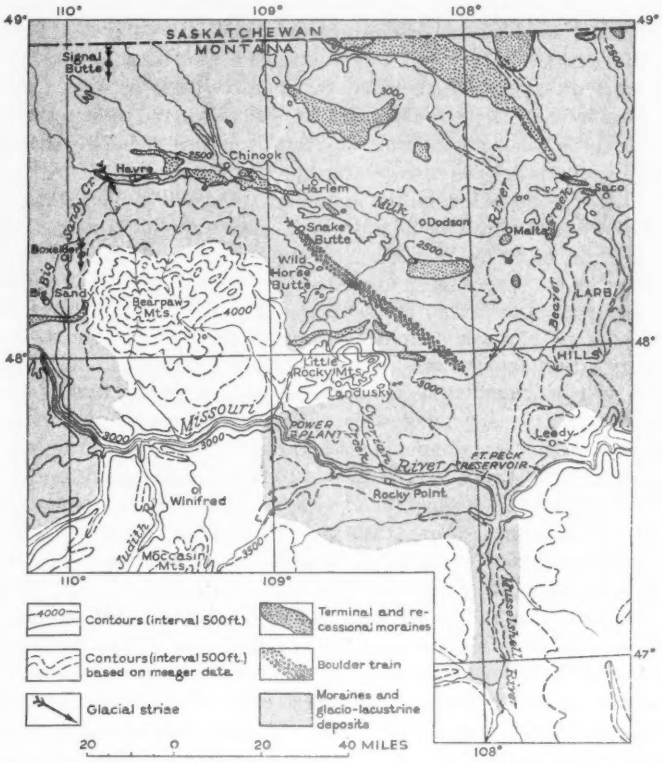

FIGURE 4.—*Same area as Figure 2, showing present configuration and glacial phenomena*

locally immobilized by included rock. The bottom of a deep glacier must therefore tend to move horizontally in directions of diminishing pressure. Apparently in accordance with this principle the rock debris in the basal part of the Keewatin glacier was transported from the locality north of the Bearpaw Mountains, where the glacier seems to have been thickest and the pressure in its basal part therefore greatest, toward areas of less thickness and less pressure in the vicinity of the Little Rocky Mountains.

The approximate directions of travel of glacial erratics are shown by arrows drawn normal to the form lines (Fig. 3). When such motion was arrested, owing probably to glacial wastage, the sole of the Keewatin glacier was carrying a train of shonkinite boulders that extended at least 49 miles southeastward from the Snake Butte intrusive. When the glacier disintegrated the train of boulders was deposited with the ground moraine. The shonkinite boulders near Midale, which are farthest from the intrusive, were presumably the first blocks of this type to be picked up, but it is possible that blocks of the less conspicuous syenitic rock of the upper part of the intrusive were removed from it earlier and may have been carried farther. A further search for this syenitic rock in the drift southeast of Midale might extend the known length of the boulder train.

As the Snake Butte boulder train is approximately straight and is narrow throughout its known length, the ice current that conveyed the boulders did not spread much and must have maintained a constant direction, for if the direction of motion had changed during transportation of the boulders, they would probably have spread out to form a fanlike boulder train of the type most commonly described in the literature.

The zone of ice moving southeastward past the Snake Butte intrusive must have been many miles wide. The straight course of the boulder train to a point near Midale, beyond the Little Rocky Mountains, shows that the southeastward motion in this zone was undeflected on the west side of the preglacial Musselshell valley. It may have maintained its direction some distance farther and if it persisted for 15 or 20 miles beyond the most southeasterly shonkinite boulder yet found, the current of ice must have encountered the west side of the Larb Hills upland (Fig. 2).

Glacial drift on the highest parts of the Larb Hills (Fig. 4) indicates that some ice overrode them and that the ice in the valley between them and the Little Rocky Mountains was many hundreds of feet thick. How far south the ice advanced in the Musselshell valley into the area shown as a marginal lake on Figure 3 has not been determined as it is not known in how much of this valley the glacial debris was deposited solely by the ice floating in the marginal lake. Nevertheless some rock debris must have been carried southward by the glacier and the ice margin may have advanced as far south as the present Missouri River. A small lobe, extending a few miles southwestward, deposited the drift mapped on Figure 4 in the saddle between the Bearpaw and Little Rocky Mountain masses. The terminal moraine mapped by Alden (Fig. 4) in the plains along the north side of the Little Rocky Mountains was deposited by ice moving into this small lobe,—representing a lateral flow from the mass that moved southeastward past Snake Butte.

It is assumed that the glacier behaved similarly during each advance,

but the Musselshell lobe may have developed only during the last advance, when the Snake Butte boulder train is believed to have formed, or only in an earlier advance, perhaps of Iowan or Illinoian age, somewhat as indicated by Alden (1932, p. 35).

Rutherford (1941) describes large quartzite boulders in the Rocky Mountain foothills belt of western Alberta, occurring in the drift from the Keewatin ice for distances up to 125 miles from their source. These erratics, like the shonkinite boulders southeast of Snake Butte, travelled parallel to the Keewatin drift margin, the elevation of which in both localities decreases in the direction of travel. Rutherford concluded that the quartzite boulders were transported chiefly by ice floating in bodies of water impounded in front of the retreating glacier. This is identical with Alden's (1932, p. 107) suggestion regarding the manner in which the Snake Butte boulders were transported, whereas in the present study these boulders are regarded as having moved within the glacier itself along a line nearly parallel with its margin.

THICKNESS OF THE ICE

The thickness of the ice mass that covered the plains of the Fort Belknap Indian Reservation is described in the following words by Alden (1932, p. 100):

> ". . . the higher of the Twin Buttes rises to an altitude of 3,850 feet, or about 850 feet higher than the plain at its base and the moraine at its most southerly point. Inasmuch as drift pebbles were found on the top of this butte, it appears that the ice must have been at least 850 feet thick within 12 miles of its limit. This butte is about 7 miles farther north and 350 feet higher than the limit of the ice advance on the high benches north of Lodgepole. The top of this butte is about 1,500 feet higher than the flat along the Milk River 20 to 25 miles to the north and northeast. Probably there was a considerable thickness of ice even on top of the Twin Buttes, but if it is supposed that the highest point was barely covered and if a southward slope of only 20 feet to the mile is assumed for the surface of the glacier, there must have been about 1,000 feet of ice on top of Snake Butte and at least 2,000 feet of ice over the flats along Milk River."

Glacial erratics at altitudes above 5000 feet in the Bearpaw Mountains indicate a thickness of at least 2500 feet in the valley of Milk River near Havre.

PERMANENT EFFECTS UPON DRAINAGE

The preglacial course of the Missouri River, according to Calhoun (1906, p. 34-45) and Alden (1932, p. 89-93), extended northeastward along the valley of Big Sandy Creek, a tributary of Milk River west of the Bearpaw Mountains, and it followed the valley of Milk River east of Havre (Fig. 2). The Musselshell River flowed northward along the valley of Beaver Creek between the Little Rocky Mountains and the Larb Hills joining the preglacial Missouri near Saco. A preglacial tributary of the Mussellshell, heading in the Bearpaw and Moccasin Mountains, followed the course of the present Missouri eastward between the

Judith Mountains and the Little Rocky Mountains. A divide separated the respective basins occupied by this tributary and by the Judith River. A continuous highland extended from the Larb Hills southward on the east side of the Musselshell River.

When the preglacial Missouri was blocked by ice west of the Bearpaw Mountains its course was diverted eastward into the basin, south of the Bearpaw and Little Rocky Mountains, through which the river flows today. The Musselshell was intercepted at its present mouth and a gorge was cut by the Missouri through the preglacial Larb Hills upland near Leedy.

Calhoun (1906, p. 40-45) concluded that the Shonkin Sag, on the north side of the Highwood Mountains, was the outlet for water from Great Falls Lake while the ice front stood near the base of the Highwood Mountains. He refers (1906, p. 44) only briefly to "the lake that existed while the ice was advancing from the former mouth of the Missouri near Havre". As this temporary lake was fed at first by all of the drainage water from the basins of the Judith, Arrow, Missouri, Sun, Teton and Marias rivers, the lake must have grown to large proportions in a relatively short time and must have attained a considerable depth before the ice front had advanced as far south as Fort Benton. If the water surface rose to an altitude between 3000 and 3200 feet, its height, judging from data given by Reeves (1927, p. 73), would have equalled that of the lowest part of the broad preglacial saddle between the Great Falls and Musselshell basins. Rising still higher it would have spilled over the saddle eastward into the basin of the Musselshell River, cutting a deep channel in the soft Cretaceous sedimentary rock which forms the saddle. If at this time the Keewatin ice blocked the preglacial Musselshell valley, between the Larb Hills and the Little Rocky Mountains, the water from Great Falls Lake, added to that of the Musselshell and its tributaries, would have rapidly filled the Musselshell basin to capacity. Here again, if the lake level attained an altitude between 3000 and 3200 feet, water overflowing into the valley of Big Dry Creek, the next basin to the east, would have cut a channel through the preglacial Larb Hills. The origin of the gorge, cut several hundred feet deep in soft shale, through which the Missouri River now flows east of Leedy can be satisfactorily explained in this manner. According to Calhoun (1906, p. 45), "the same process was repeated in the valley of the Big Dry" and the present course of the Missouri was eventually established from the Great Falls basin to the mouth of Milk River, east of Fort Peck.

MAXIMUM EXTENT OF THE ICE

The entire drift-covered area in the plains adjacent to the Little Rocky and Bearpaw Mountains may have been covered by the ice of the Keewatin glacier at the time of its maximum advance, but the evidence for its extension south of the Missouri River is not altogether conclusive. Reeves (1924, p. 72; 1927, p. 42) suggested that the drift south of the river might have been rafted into position by floating ice, and Alden's statement in this connection (1932, p. 81, 82) follows:

> ". . . some, at least, of the scattered pebbles and boulders may have been transported by ice floating on water ponded in front of the ice sheet as it lay along or near the present course of the Missouri River and . . . there may have been an outlet for this lake southeastward from the Musselshell Valley to the Yellowstone near Forsyth. There must have been a great deal of water flowing eastward or southeastward somewhere through this region for a time, for all the drainage of that part of the Missouri Basin outside the margin of the ice sheet, together with water from the glacier itself, must have had an outlet to the Yellowstone. The absence of any known glacial till south of the Missouri River within this tract makes it somewhat questionable whether or not the ice lobe between the Little Rocky Mountains and the Larb Hills on the east extended southward beyond the present line of the Missouri. It may be noted, however, that neither have lacustrine clays, such as are present in the region of Great Falls, been observed in the tract described above. There is no question that the ice between the Little Rocky Mountains and the Larb Hills advanced nearly if not quite to that part of the Missouri's present course here under discussion."

That the ice margin did advance at least as far as the Missouri River is suggested by the presence of till-like material at the mouth of Cyprian Creek (Fig. 4) and at Rocky Point; and glacial material in the Musselshell valley and on the north bank of the Missouri opposite the mouth of the Musselshell is mentioned by Calhoun (1906, p. 55) with the statement that "it may have been deposited either by ice or by water"; the nature of this material is not further described. Moreover, glacial pebbles and boulders are abundant along the road from a point northeast of Winifred to the crossing of the Missouri River at the Power Plant ferry. If they are as numerous in the rest of the drift-bearing area indicated by Reeves, there is reason to suspect that the ice sheet once extended into the country south of the Missouri River. Absence of a terminal moraine in that part of the Musselshell lobe does not necessarily indicate that the ice sheet was never present in the area south of the Missouri, but suggests that the drift in this area may have been deposited during a brief period of active glacial expansion, followed either by immediate disappearance or by stagnation of the ice. It is even possible that the apparently driftless area included in the Winifred re-entrant was partly covered by ice.

The ice sheet probably impounded a large amount of water, temporarily, when its margin extended across the part of the pre-glacial

valley of the Musselshell that is now occupied by Beaver Creek (Fig. 4), but if the water of a marginal lake covering the highest parts of the large area occupied by the drift south of the Little Rocky Mountains remained long enough to permit any large proportion of the abundant drift that is present to accumulate by the slow-ice-rafting process, this lake would be expected to leave unmistakable signs of its existence, such as ancient shore lines or lacustrine sediments. Nothing of the kind has been recognized. The quantity of drift that is present is as great as in many areas mapped as "ground moraine" farther north and may be largely the result of extension of a lobe of the Keewatin glacier into the drift-covered area south of the Little Rocky Mountains. The Musselshell lobe is indicated on Calhoun's map (1906, Pl. 1) by an area of "older drift" extending southwestward across the Missouri River.

RETREAT

The Winifred re-entrant (Pl. 1) probably persisted, increasing gradually in size, as the ice melted back northward. The broad arcuate recessional moraines (Fig. 4) extending northwestward from Saco to the Canadian border and from the Larb Hills along Milk River to Havre suggest a pattern resulting from the growth of this re-entrant. Many other interesting recessional phenomena of this part of Montana are described by Alden (1932, p. 110-114) but do not fall within the scope of the present discussion. The question of differential uplift of the Great Plains during or following retreat of the Keewatin ice is discussed by Alden (1932, p. 130) and Rutherford (1941, p. 115-118).

REFERENCES CITED

Alden, W. C. (1932) *Physiography and glacial geology of eastern Montana and adjacent areas,* U. S. Geol. Survey, Prof. Paper 174.

Antevs, E. (1922) *The recession of the last ice sheet in New England,* Am. Geogr. Res., ser. no. 11, New York.

Buell, I. M. (1895) *Boulder trains from the outcrops of the Waterloo quartzite area,* Wis. Acad. Sci., Tr., vol. 9, p. 255-274.

Calhoun, F. H. H. (1906) *The Montana lobe of the Keewatin ice sheet,* U. S. Geol. Survey, Prof. Paper 50.

Cathcart, S. H. (1922 ?) U. S. Geol. Survey, unpublished manuscript.

Collier, A. J., and Thom, W. T., Jr. (1918) *The Flaxville gravel and its relation to other terrace gravels of the northern Great Plains,* U. S. Geol. Survey, Prof. Paper 108-J, p. 179-184.

Dobbin, C. E., and Erdmann, C. E. (1934) *Geology and occurrence of petroleum in the United States; Petroleum investigation,* pt. 2 (U. S., 73d Cong., H. R., Committee on Interstate and Foreign Commerce, Hearings before a subcommittee on H. Res. 441), p. 1053, fig. 120.

Fraser, F. J., *et al.* (1935) *Geology of southern Saskatchewan,* Canada Dept. Mines, Geol. Survey, Mem. 176, Publ. 2373.

Hausen, H. (1912) *Sydfinska ledblockens spridning i Ryssland,* Comm. Géol. Finlande, Bull. 32.

Hess, H. (1933) *Das Eis der Erde,* Handbuch der Geophysik, Band VII, p. 1-121, Berlin.

Hopkins, O. B. (1923) *Some structural features of the plains area of Alberta caused by Pleistocene glaciation,* Geol. Soc. Am., Bull. 34, p. 419-430.

Hurlbut, C. S., Jr. (1939) *Igneous rocks of the Highwood Mountains, Montana,* Pt. 1, *The laccoliths;* with section on structure and mechanism of intrusion by David T. Griggs, Geol. Soc. Am., Bull. 50, p. 1063-1065.

McConnell, R. G. (1885) *Report on the Cypress Hills, Wood Mountain, and adjacent country,* Canada Geol. Survey, Ann. Rept. no. 1.

Pecora, W. T. (1941) *Structure and petrology of the Boxelder laccolith, Bearpaw Mountains, Montana,* Geol. Soc. Am., Bull. 52, p. 817-854.

Reeves, F. (1924) *Geology and possible oil and gas resources of the faulted area south of the Bearpaw Mountains, Montana,* U. S. Geol. Survey, Bull. 751-C.

——— (1927) *Geology of the Cat Creek and Devils Basin oil fields and adjacent areas in Montana,* U. S. Geol. Survey, Bull. 786-B.

Rutherford, R. L. (1928) *Two interesting boulders in the glacial deposits of Alberta,* Jour. Geol., vol. 36, no. 6.

——— (1941) *Some aspects of glaciation in southwestern Alberta,* Royal Soc. Canada, Tr., Sec. 4. vol. 35.

Sauramo, M. (1929) *The Quaternary geology of Finland,* Comm. Géol. Finlande, Bull. 86.

Shaler, N. S. (1893) *The conditions of erosion beneath deep glaciers, based on a study of the boulder train from Iron Hill, Cumberland, R. I.,* Harvard Coll., Mus. Comp. Zool., Bull., vol. 16, p. 185.

Taylor, F. B. (1910) *The Richmond and Great Barrington boulder trains,* Geol. Soc. Am., Bull. 21, p. 747-752.

Tyrrell, J. B. (1898) *The glaciation of north central Canada,* Jour. Geol., vol. 6, p. 147-160.

Weed, W. H., and Pirsson, L. V. (1895) *Highwood Mountains of Montana,* Geol. Soc. Am., Bull. 6, p. 389-422.

Woodworth, J. B., *et al.* (1935) *Geography and geology of the region including Cape Cod, the Elizabeth Islands, Nantucket, Marthas Vineyard, No Mans Land, and Block Island,* Harvard Coll., Mus. Comp. Zool., Mem., vol. 52.

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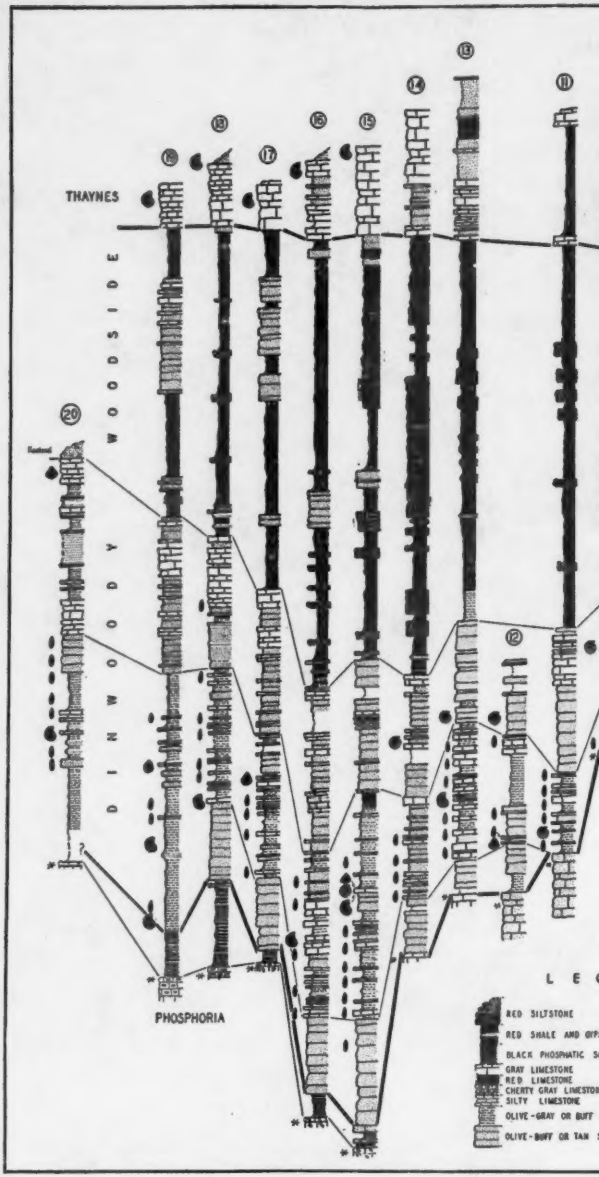

CORRELATION OF OUTCROP SECTIONS OF
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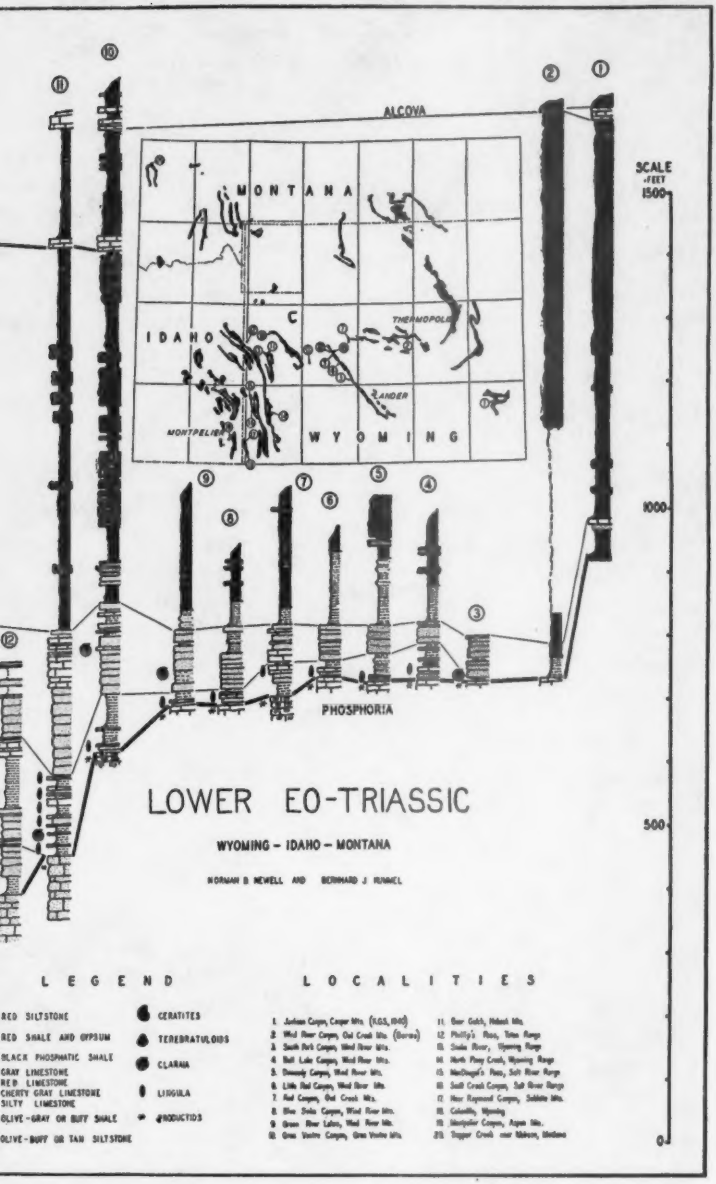

ONS OF DINWOODY AND WOODSIDE FORMATIONS IN
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LOWER EO-TRIASSIC STRATIGRAPHY, WESTERN WYOMING AND SOUTHEAST IDAHO

BY NORMAN D. NEWELL AND BERNHARD KUMMEL

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ABSTRACT

A thickness of from 1000 to 2000 feet of strata in western Wyoming, southeastern Idaho, and southwestern Montana is shown to belong to the *Otoceras,* and probably to the *Genodiscus,* zones of the Scythian stage of the Lower Triassic. These strata,

herein classified as the Dinwoody and Woodside formations, exhibit marked overlap relations on the underlying Phosphoria formation, indicative of a marked hiatus corresponding to the Permo-Triassic boundary. Unlike current usage in southeastern Idaho the term Woodside is herein employed for unfossiliferous red beds, which apparently grade northeastward into a nonred fossiliferous section similar to the underlying Dinwoody formation. Further work remains to be done in the areas in which the red-beds facies is lacking. The marine invertebrate fauna of the Dinwoody has a marked affinity with the *Otoceras* fauna from east Greenland, Seis (lower Werfen) fauna of the Alps, and early Triassic faunas from eastern Siberia.

INTRODUCTION

Throughout a vast area in southwestern Montana, southeastern Idaho, western Wyoming, Utah, northern Arizona, and southern Nevada the marine limestones (Phosphoria and Kaibab) of the Permian are overlain by a variable thickness of fine, calcareous, greenish-buff to gray marine clastics, with local red-beds tongues. In the northern part of the area these beds constitute the Dinwoody and Woodside formations. In the southern part of the area they commonly are classed in the Moenkopi formation. Save for a very few notable exceptions the shales, thin limestones, and siltstones appear to be conformable on the Permian limestones. It is almost certain, however, that a hiatus of considerable magnitude occurs at the top of these limestones throughout much of western and northwestern United States. An erosional disconformity has been recognized locally between the Kaibab and Moenkopi in Arizona and Nevada (McKee, 1938, p. 54). According to J. Stewart Williams (personal communication) the Moenkopi formation rests disconformably on the Kaibab limestone along the Hurricane Cliffs in southern Utah. In this area the basal beds (Rock Canyon conglomerate) of the Moenkopi lie on eroded beds of limestone in the Kaibab, and, more significantly, the limestones 15 feet above the Kaibab are lithologically like the Thaynes formation of northern Utah and contain the lower Thaynes *Meekoceras* fauna. Since the base of the *Meekoceras* beds is more than 1000 feet above the Permian limestones in northern Utah a marked convergence to the southward is indicated between the two horizons. Whether or not there is also a southward overlap of the Moenkopi on the Kaibab across Utah has not been determined. According to Baker and Williams (1940, p. 624) Permian limestones are overlain nonconformably by the Woodside shales in the region just east of Provo, in northern Utah. Here the Woodside bevels across nearly 2000 feet of Permian strata in a horizontal distance of about 10 miles. Farther northeastward, in western Wyoming, physical evidence for an unconformity at the base of the Dinwoody formation is recognized in (1) marked leaching of cherty beds at the top of the Phosphoria formation, (2) marked northeastward overlap of lower Dinwoody strata by upper Dinwoody on the Phosphoria surface from southwestern Wyo-

ming into central Wyoming, and (3) discovery by Alfred Fischer (personal communication), during current stratigraphic studies on the Phosphoria formation, that upper beds of the Phosphoria are locally truncated by basal beds of the Dinwoody formation. In the same area Love (1939, p. 33, 40) also has noted local evidences of unconformity at the boundary between the Phosphoria and Dinwoody formations. It is well to stress, however, the lack of convincing physical evidence for a hiatus at most exposures throughout Wyoming and southeastern Idaho.

According to faunal evidence the youngest Permian recognized within the Middle Rocky Mountain area and the Colorado Plateau area belongs to the Leonard series and to the lower Guadalupian (Word) (King, 1931, p. 31-32; Longwell and Dunbar, 1936, p. 1199-1202; Love, 1939, p. 31-38; Miller and Cline, 1934, p. 284; Miller and Furnish, 1940, p. 23). The distinctive guide fossils, however, on which the age determinations have been based, generally were not collected from the uppermost part of the Permian limestones. Furthermore, the faunas of the overlying Dinwoody, Woodside, and Moenkopi formations were virtually unknown. Consequently it is not surprising that several geologists (Thomas, 1934, p. 1669; Love, 1939, p. 41) have recognized a possibility that part or all of the Dinwoody is Permian. It was with the hope of locating the boundary between the Permian and Triassic systems in the western part of the Middle Rockies that the writers undertook the present investigation. It is possible that post-Word rocks occur in the upper part of the Phosphoria (Rex chert, post-Rex limestones and shales), but findings indicate definitely an early Triassic age for all the strata in the Dinwoody and Woodside formations.

When first described in the Park City district, near Salt Lake City, the rocks currently classed as Woodside and the overlying Thaynes were termed "Permo-Carboniferous." Later Boutwell (1912, p. 42-59) correlated the Thaynes with the Lower Triassic *Meekoceras* zone of Hyatt and Smith (1905, p. 17-19). Because the Woodside seemed lithologically more closely allied to the Thaynes than to the underlying Paleozoic rocks Boutwell and the majority of later workers tentatively regarded the Woodside as probably also of early Triassic age. In his classical treatise on the geology of southeastern Idaho, Mansfield (1927, p. 84, 85) wrote that "the relations of the Triassic to the Permian in the region are not definitely known." However, influenced principally by lithologic criteria, together with scanty faunal evidence furnished by G. H. Girty, he concluded that the Woodside probably belongs to the Lower Triassic rather than to the Permian system.

The type locality of the Woodside formation is in the Park City district, Utah, not far from Salt Lake City. In this area the Woodside over-

lies the Phosphoria formation (J. S. Williams, 1939, p. 82-100). In the Wind River Mountains near Lander, Wyoming, 200 miles to the northeast, the Phosphoria formation is overlain by 40-120 feet of drab shales and siltstones, known as the Dinwoody formation.

There has been considerable confusion regarding the stratigraphic relationship of the Woodside and Dinwoody. Blackwelder (1918, p. 425-426) thought that the Dinwoody, a relatively thin formation, is approximately equivalent to the Woodside and Thaynes formations, which aggregate more than a mile of strata in southeastern Idaho. Others (Love, 1939, p. 41) have supposed that the Dinwoody includes equivalents of part or all of the Woodside. Present findings indicate that the Dinwoody is the stratigraphic equivalent of approximately the lower half of the Woodside as employed by Mansfield (1927, p. 86) for southeastern Idaho. However, as will be shown, it is undesirable to employ the term Woodside in the sense in which Mansfield used it.

In a preliminary notice the writers (Newell and Kummel, 1941, p. 204-208) offered evidence in support of the suggestion by Muller and Ferguson (1939, p. 1583) that the strata between the Phosphoria and the Thaynes formations might contain the *Otoceras* and *Genodiscus* zones of the Eo-Triassic series. After a second field season the writers present further faunal evidence in support of the preliminary conclusions and describe a stratigraphic overlap at the base of the Dinwoody formation.

Kummel is continuing the studies on Lower Triassic stratigraphy of the southeast Idaho region and is expanding the work to include the Thaynes formation.

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STRATIGRAPHY

DINWOODY FORMATION

General considerations.—The Dinwoody formation was named and defined by Blackwelder (1918, p. 425-426) from outcrops in Dinwoody

Canyon on the northeastern slope of the Wind River Mountains, near Du Bois, Wyoming. As originally defined the formation comprised 200 feet (the writers' measurement) of gray and olive shaly siltstones and shales, with thin brown limestones near the base. The limits of the formation were defined by the Phosphoria limestone, below, and the bright red shales and siltstones of the Chugwater formation, above. The writers' research reveals that the change from gray to red beds does not follow a stratigraphic plane but varies in position from place to place, even within a few hundred yards in Dinwoody Canyon. Because the color boundary between the Dinwoody, as originally defined, and the overlying red Chugwater is not a useful nor natural boundary, it is proposed to restrict the Dinwoody at the type locality to include only dominantly silty strata between the Phosphoria and the top of resistant siltstones about half way toward the summit of the original Dinwoody. As redefined, the Dinwoody is 90 feet thick at the type locality. The resistant siltstones at the top are overlain abruptly by soft shales, which generally are red, but locally are gray, as at Dinwoody Canyon.

Throughout the Wind River Mountains and the western part of the Owl Creek Mountains, west-central Wyoming, the formation is relatively thin, ranging from less than 40 feet southeast of Lander to around 120 feet at the northwest end of the Wind River Mountains. South and southwest of the Jackson Hole area, western Wyoming, the Dinwoody thickens progressively, attaining a thickness of around 650-700 feet in the deeper parts of the geosyncline near Afton and Cokeville, Wyoming, and Montpelier, Idaho. Some of the increase toward the southwest is produced by the addition of new beds at the base of the formation, but there is also a general increase in thickness in the more persistent subdivisions. Dense, blocky, thin limestones appear in the formation where it is relatively thick and constitute a characteristic feature of the unit.

Three major divisions can usefully be recognized in the Dinwoody. Although they are lithologically distinctive, and relatively extensive, the writers have chosen to avoid the application of new geographic names to the divisions until further work in Idaho has tested the usefulness of this classification. For the present report the lowest division, geographically limited to the eastern flank of the geosynclinal area, is called the *basal siltstone*. The second unit, characterized by a large number of thin, dense limestones and dark gray to olive shales, both of which commonly contain very abundant and well-preserved *Lingula borealis* Bittner, is designated the *Lingula* zone. The upper division consists principally of calcareous, resistant olive-buff to brown siltstones, with some interbedded hard, gray limestones. In the southwestern areas, where the unit is relatively thick, massive silty limestones appear in the upper part. Because

of the abundance of the distinctive pelecypod *Claraia stachei* Bittner, especially in the less limey facies of the northeastern outcrops, this unit is provisionally called the *Claraia* zone.

Basal siltstone.—The basal deposits of the Dinwoody in the Teton, Wyoming, and Salt River ranges (Pl. 1, sections 12-18) consist of friable, unfossiliferous, buff siltstone, ranging from about 50 feet to around 175 feet, being thickest in the southern exposures. In the areas east of the Wyoming Range (Pl. 1, sections 1-11) the basal siltstone unit is absent and apparently is overlapped by the *Lingula* beds, which extend farther toward the east. At Green River Lakes (Pl. 1, section 9) and another locality near the head of Twin Creek, in sec. 22, T.38N., R.111W., the basal deposit of the Dinwoody is a 4-foot bed of medium-grained brown sandstone, which might be the stratigraphic equivalent of the basal siltstone of the western area. Sandstone in the Dinwoody is quite exceptional, however, and correlation of these isolated occurrences is conjectural.

The basal siltstone has not thus far been recognized in southeastern Idaho. The lower part of the Dinwoody is there represented by gray and buff silty shale, which probably is the equivalent of the siltstone of southwestern Wyoming.

Lingula zone.—Overlying the basal siltstone in western Wyoming, or resting on the Phosphoria formation in the western part of the Owl Creek and Wind River Mountains (Pl. 1, sections 4-9), is a somewhat heterogeneous unit which is especially distinguished by the profusion, at least locally, of exquisitely preserved *Lingula*. Although the Lingulas occur sparsely both above and below the *Lingula* zone, as here defined, they are nowhere else abundant. The beds of this zone consist dominantly of silty olive-buff to gray shales, interbedded with thin, blocky limestones, ranging from brown to dark bluish-gray. In general the limestones are silty and in a very few instances might be mistaken for siltstones on weathered surfaces. More commonly, however, the limestones do not change color appreciably on weathering. Variable, more or less calcareous siltstones occur throughout the section, probably representing lateral equivalents of the silty shales. The *Lingula* zone reaches a maximum observed thickness in the vicinity of Afton, Wyoming (Pl. 1, sections 15 and 16), where it ranges in thickness from about 250 to 350 feet. Toward the north and east it thins progressively and comes to rest directly on the Phosphoria formation in the Hoback, Gros Ventre, Owl Creek, and Wind River Mountains (Pl. 1, sections 4-11). Throughout the eastern area of its outcrop the *Lingula* zone is thin, commonly measuring from 25 to 50 feet. To the east of the longitude of Lander, Wyoming, the *Lingula* beds appear

to be overlapped, because the *Claraia* zone rests on the Phosphoria formation in the southeastern part of the Wind River Mountains, as for instance at Hidden Anticline, 12 miles southeast of Lander. The fauna of the *Lingula* zone, in addition to the great numbers of Lingulas, consists of a few species of pelecypods, a brachiopod or so, and rare, poorly preserved ceratites of the *Ophiceras* type.

Claraia zone.—The uppermost and most extensive unit of the Dinwoody consists dominantly of resistant, light-brown, calcareous siltstones, generally bedded in even layers less than a foot thick. Locally it contains interbedded light gray, silty limestones, which range up to more than 50 feet in thickness in the deeper parts of the geosyncline. The strata of this zone are distinguished lithologically from the preceding zone chiefly by the relatively small content of shale. It is highly probable that the majority of the limestone beds in the *Claraia* zone grade laterally into calcareous siltstone beds within short distances, because none of the limestones is known to be very persistent. In general the limestones are lighter in color, more massive, less fossiliferous, and somewhat more pure than the limestones of the *Lingula* zone. In the southeastern part of the Wind River Mountains, as at Hidden Anticline, 12 miles southeast of Lander, Wyoming, the beds of the *Claraia* zone lie directly on the Phosphoria formation, and the lower units of the Dinwoody are missing, apparently by nondeposition, so that the eastward transgressive overlap of the formation is particularly striking in this area. In the Wind River and Owl Creek Mountains the beds of the *Claraia* zone range in thickness from 25 or 30 feet to nearly 100 feet (Pl. 1, sections 3-9). In the mountains of western Wyoming the unit increases in thickness to around 300 feet and becomes relatively more calcareous.

The silty beds of the *Claraia* zone are locally replete with molds of a few kinds of pelecypods, of which species of *Claraia* are the most distinctive. A few terebratuloid and spiriferoid brachiopods occur in the purer limestones but are exceedingly hard to collect.

In the areas to the east of the Owl Creek and Wind River Mountains equivalents of the Dinwoody formation (Pl. 1, sections 1 and 2) change to red silty shales, except for a buff calcareous and gypsiferous zone near the top. Thomas (1934, p. 1669) has given the name Little Medicine to these nonred rocks. It seems likely that they correspond to the upper part of the *Claraia* zone.

Nonred calcareous and gypsiferous siltstones occur above the Phosphoria at various places in the Bighorn Basin and at Cinnabar Mountain, at the northern entrance to Yellowstone Park, locality 23. The writers' measurement of 63 feet for the Dinwoody equivalents at Devil's

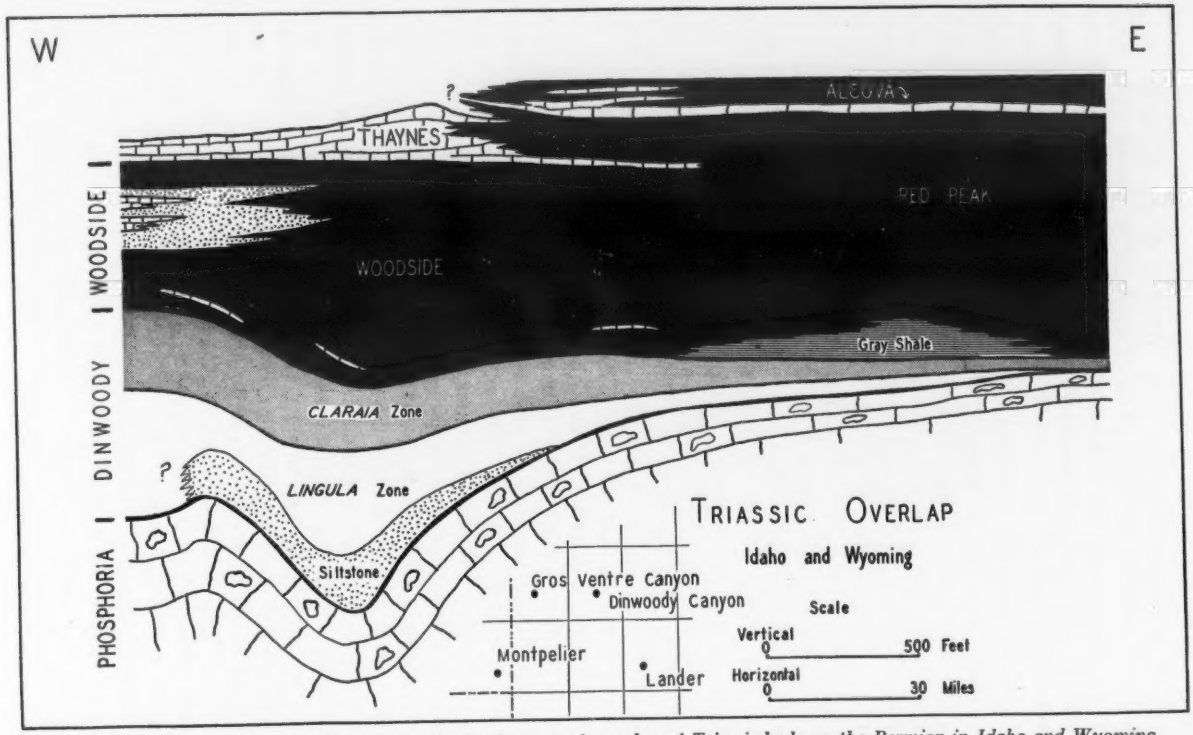

FIGURE 1.—*Schematic diagram illustrating the northeastward overlap of Triassic beds on the Permian in Idaho and Wyoming*

Slide, Cinnabar Mountain, does not agree with Charles W. Wilson's (1934, p. 372) measurement of 35 feet. These rocks appear lithologically referable to the *Claraia* zone, although the Dinwoody generally is not fossiliferous in these areas.

Dinwoody overlap.—The successively higher divisions of the Dinwoody formation are progressively more extensive toward the north and east across southwestern Wyoming. The basal siltstone generally is not recognizable east of the Wyoming and Teton ranges. In a general way the eastern margin of this unit, corresponding as it does with the east front of the area of overthrust faults of southwestern Wyoming, probably delineates the approximate border of the Lower Triassic geosyncline. Toward the east the basal siltstone is missing, probably because of nondeposition. As the geosyncline was filled by sediments the rate of filling tended to exceed the rate of subsidence, so that beds of the *Lingula* zone were spread more widely toward the east than the lower beds and reached at least as far as the longitude of Lander, Wyoming, in the Wind River Mountains. Beds of the *Claraia* zone are spread even farther toward the east, an unknown distance beyond Lander, suggesting that subsidence along the geosynclinal axis was checked or at least was not so localized as previously. If the upper part of the *Claraia* zone corresponds to Thomas' (1934, p. 1669) Little Medicine tongue the equivalents of the *Claraia* zone can be recognized as far to the east as southeastern Wyoming.

WOODSIDE FORMATION

Boutwell (1907, p. 446) named the Woodside formation from Woodside Canyon in the Park City mining district, northeastern Utah. In the type area the formation consists of a little over 1000 feet of maroon and red shaly siltstone, barren of fossils. It is bounded below by the Phosphoria formation (J. Stewart Williams, 1939, p. 96) and above by the Thaynes formation. In southeastern Idaho it is usually not red, and the interval between the Phosphoria and Thaynes—1000 to 2000 feet and more—is grayish-brown and drab silty shales, siltstones, and minor amounts of bluish-gray limestones. In the vicinity of Montpelier the upper third of this interval contains conspicuous red and maroon rocks. Reconnaissance observations by various geologists, confirmed in part, indicate that the red beds grade into fossiliferous nonred rocks a short distance north of Montpelier, because no trace of red shales and siltstones was observed at the Blackfoot Reservoir, near Henry, Idaho (locality 21) where there is more than 2000 feet of nonred rocks of Dinwoody lithology between the Phosphoria and Thaynes. It seems evident that the nonred equivalents of the Woodside in the country north of Montpelier, Idaho, belong

FIGURE 1.—*Schematic diagram illustrating the northeastward overlap of Triassic beds on the*

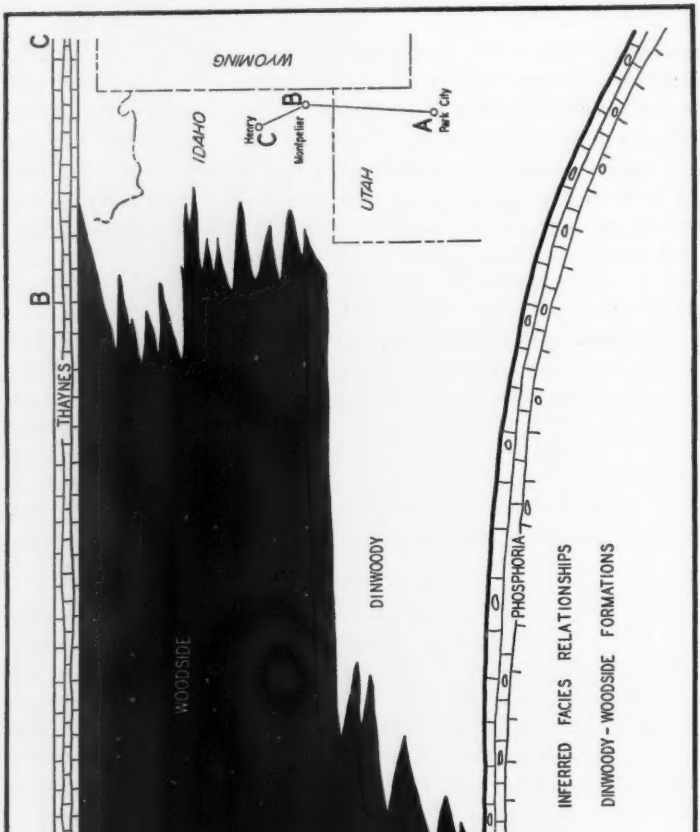

FIGURE 2.—*Inferred facies changes in the Dinwoody and Woodside formations between Park City, Utah, and Henry, Idaho*

to quite a different lithologic facies and deserve a separate name. However, understanding of this gray facies of the Woodside is not yet sufficiently complete to justify erection of a separate formational unit.

The Dinwoody formation is probably the time equivalent of the lower part of the red-beds section in the type area of the Woodside, and it might, therefore, be argued that the Dinwoody should be regarded as a subdivision, a member, of the Woodside formation. This seems wholly inconsistent, however, with the general understanding of a formation as a lithologic and genetic unit—to class the fossiliferous nonred units and the unfossiliferous redbeds in a single formation.

In much of western Wyoming the Dinwoody is overlain by several hundred feet of red shaly siltstones and sandstones, variously called Lower Chugwater, Red Peak, and "red Woodside". These red beds belong to the same facies as the type Woodside but apparently correspond stratigraphically only to about the upper half of the typical Woodside as defined in the Park City area. A conservative procedure is to regard these red beds between the Dinwoody and the Thaynes as a northeastern tongue of the Woodside. When the relations with the type Woodside are better known it may prove to be desirable to apply a formal name to this tongue of the Woodside. The following discussion is concerned principally with the red-beds facies between the Dinwoody and Thaynes. It is obvious that a very important field for further investigation lies in the area north of Montpelier, Idaho, where the red-beds facies is missing.

FIGURE 2.—*Inferred facies changes in the Dinwoody and Woodside formations between Park City, Utah, and Henry, Idaho*

WOODSIDE-THAYNES BOUNDARY

Throughout southeastern Idaho and the southern part of the Salt River range in southwestern Wyoming the basal unit of the Thaynes is a massive, cliff-forming gray and white limestone up to 150 feet in thickness. An abundance of ceratites of the *Meekoceras* zone occurs in at least one horizon wherever the lower Thaynes in this region was examined. Commonly the cephalopods are packed together in great numbers through an interval not greater than about 20 feet, and elsewhere in the limestone they are very rare. The cephalopod bed occurs some distance above the base of the Thaynes (Pl. 1, sections 15-19), in some instances at the extreme top of the basal limestone 100 feet and more above the red siltstones of the Woodside. In other localities the position of abundant cephalopods is apparently near the middle of the basal limestone. Further work, now in progress by Kummel, may reveal that the *Meekoceras* fauna occurs at more than one stratigraphic level.

The lower Thaynes is lithologically well defined and easily recognized as far northeastward as the canyon of Snake River in the Wyoming range, near Hoback post office in western Wyoming (locality 13). Reconnaissance observations indicate profound changes in the Thaynes in the score or so miles between this locality (Pl. 1, sections 13, 11, 10) and sections in Hoback Canyon and Gros Ventre Canyon just east of Jackson Hole. Although time did not permit more than cursory examination of the Thaynes equivalents there appears to be a profound convergence or overlap of the Thaynes toward the north and east. At the locality 13, near Hoback post office, on Snake River, the Thaynes includes probably 1000 feet or more of silty limestones and buff shales, with only minor quantities of red beds. Thaynes equivalents in Hoback Canyon and Gros Ventre Canyon are apparently less than half as thick and consist principally of red beds with relatively inconspicuous purplish to white dolomitic limestones. Some of these limestones are lithologically like the persistent Alcova limestone of eastern Wyoming and the Wind River Mountains. On stratigraphic position one of the upper limestones in Gros Ventre Canyon about 200 feet above the base of the Thaynes is correlated with the Alcova.

The Woodside-Thaynes boundary has not been recognized in the Wind River Mountains or anywhere in eastern Wyoming; consequently it seems appropriate to use a separate nomenclature for the eastern area.

RED PEAK FORMATION

Woodside equivalents in the Wind River and Owl Creek Mountains are included in the Red Peak formation (Love, 1939, p. 44), the basal unit of the red beds Chugwater terrane. In the type area, in the western Owl Creek Mountains, not far from Du Bois, Wyoming, the Red Peak is bounded above by the Crow Mountains formation, a unit of fine sandstones, which are white and gray in the Owl Creek Mountains but change to reddish siltstones in the Wind River Mountains. According to Branson and Branson (1941, p. 135) the Red Peak and Crow Mountains formations are separated by the thin but persistent Alcova dolomitic limestone in the Wind River Mountains. Inasmuch as the Alcova is the best datum horizon in the red beds of central and eastern Wyoming, it seems desirable to follow the Bransons in applying the term Red Peak to the red strata between the Dinwoody and Alcova.

The Red Peak consists of from 600 to around 800 feet of bright brick-red siltstones, some layers of which are quite shaly. The entire unit is remarkably uniform in character, and no fossils have yet been discovered in it. According to estimates based on stratigraphic position (Pl. 1, section

10) the Alcova limestone, which marks the top of the Red Peak formation, lies well above the top of the Woodside-Thaynes boundary.

PALEONTOLOGY

FAUNAL AND STRATIGRAPHIC SUMMARY

The Dinwoody fauna is of especial interest for it adds considerably to our imperfect knowledge of the earliest Triassic of North America. The fauna, although not a particularly varied one, is at least as varied

TABLE 1.—*Correlation of Lower Triassic formations in southeastern Idaho and west-central Wyoming*

Mansfield, 1927 SE Idaho	This Report		Standard faunal zones
	Western Wyoming SE Idaho	West-central Wyoming	
Thaynes	Thaynes	Crow Mountain	*Columbites*
		Alcova	*Tirolites* *Meekoceras*
Woodside	Woodside tongue	Red Peak	*Genodiscus*
	Dinwoody	Dinwoody	*Otoceras*
Phosphoria (Permian)			*Waagenoceras* (?)

and well preserved as any other contemporanous fauna thus far known, with the exception of the magnificent cephalopod fauna from East Greenland, described by Spath (1930; 1935). So little is known about the earliest Triassic faunas of the world, as well as the latest Permian faunas, that part of the apparent dissimilarities between late Paleozoic and early Mesozoic faunas is due to lack of fossil records near the boundary between rocks of the two eras. Spath has shown that basal Eo-Triassic faunas are well developed only in a few regions: the southern Alps, the Himalayas, Ussuri Bay near Vladivostok, and eastern Greenland. In these areas, as well as others where early Triassic fossils are found, the ammonites appear to be more plentiful and better known than other invertebrates. To the four important localities must now be added a fifth, the western margin of the Middle Rockies, in southeastern Idaho, western Wyoming, and southwestern Montana, where fossiliferous early Eo-Triassic rocks are thicker and more extensive than in any other part of the world.

According to Spath (1935, p. 102) the Lower Triassic, or Scythian stage, is divisible into six faunal zones:

Eo-Triassic series	Upper	Prohungaritan
		Columbitan
		Owenitan
	Lower	Flemingitan
		Gyronitan
		Otoceratan

In this scheme the Gyronitan zone is approximately equivalent to the *Genodiscus* zone of Smith (1932, p. 15), and the Flemingitan and Owenitan are included in Smith's *Meekoceras* zone. Presumably Spath's upper four zones are represented in the Thaynes formation. Preliminary examination of the Thaynes indicates the existence of horizons of cephalopods unknown to Smith. It is surprising that this finest of all Eo-Triassic sections is still poorly known. The great monographs by Hyatt and Smith (Hyatt and Smith, 1905; Smith, 1932) on Triassic ammonites of North America treat the Thaynes ammonites rather fully but do not present a single stratigraphic section to show the field relationships of their fossils! The majority of paleontologists fully realize today that detailed stratigraphic studies should go in advance of critical faunal work, but such an attitude has not always characterized paleontological work.

Ammonites were found at several localities in the Dinwoody formation. Unfortunately extensive search has thus far failed to reveal more than two identifiable specimens. All the other specimens are either flattened impressions on the laminae of shaly siltstones or are preserved in a dense matrix of limestone which does not break away from the shell. Furthermore in these latter instances sutures are either indistinguishable or are too nearly obliterated to be used for identification. The identifiable ammonites appear to be identical with *Discophiceras subkyokticum* Spath and *Metophiceras subdemissum* Spath from the upper part of the *Otoceras* zone of eastern Greenland. However, as students of Triassic faunas are well aware, primitive ceratites like *Discophiceras* and *Metophiceras* are not easy to identify, and positive identification should rest on more complete specimens than the fragmentary material now available.

In addition to the cephalopods the fauna of the Dinwoody consists of a few species of brachiopods, several pelecypod species, and one gastropod species. The majority of the forms have been described from other regions, but the chief faunal affinities appear to be with the lower Werfen beds (Seis formation) of the Alps, the Otoceratan zone of East Greenland, and the Scythian faunas of the Ussuri Bay area, eastern Siberia.

Spath (1935, p. 104-105) has shown that many of the pelecypod species range through the upper part of the Otoceratan (*Ophiceras* beds) and the lower part of the overlying Gyronitan zone (*Vishnuites* beds). Probably a positive correlation of the Dinwoody formation with one rather than the other of the two zones is not justified. However, the two ammonites suggest a late Otoceratan age.

The similarity between the writers' collections and the faunas described by Bittner (1899) and Kiparisova (1938) from the Ussuri Bay region, near Vladivostok, is very striking. Unfortunately very little is known about the stratigraphic distribution of the Ussuri fossils. Diener (1895) described a collection of cephalopods from the area and dated the beds as in the zone of *Otoceras*. Later students of cephalopods have generally assigned Diener's fauna to the Gyronitan (*Genodiscus*) zone. Kiparisova (1938, p. 301) has shown that a considerable part of the Lower Triassic section is represented in the various localities and collections from Ussuri Bay, and it is possible that some of the brachiopods and pelecypods came from horizons quite different from the one that yielded the well-known cephalopod fauna. According to Kiparisova diagnostic forms of the Flemingitan zone (lower *Meekoceras* of Smith) have been found in the region.

There are similarities between the Dinwoody fauna and a fauna from the lowermost Triassic of the Alps, the Seis formation of the lower Werfen. This fauna is generally dated as Gyronitan in terms of the ammonite zones of Spath.

A few fossils collected from the nonred equivalents of the Woodside tongue in a very thick section (over 2000 feet) near Henry, Idaho (locality 21), are identified with Siberian species of the Ussuri fauna but are not represented in the older Dinwoody collections from Wyoming. It is probable that extensive collecting from the fossiliferous beds just under the Thaynes may reveal a significant fauna. Tentatively the bulk of the Dinwoody is classed as upper Otoceratan, and the beds immediately below the *Meekoceras* zone, as Genodiscan (Gyronitan).

The earliest marine invertebrates of the Triassic were cosmopolitan, and wherever they are found they belong to a peculiar facies. Whole phyla, such as the coelenterates, echinoderms, bryozoans, protozoans, and sponges, are practically unknown in these faunas. The few brachiopods known have not been critically studied by modern techniques, but they apparently are not varied. Commonly they are referred to *Lingula*, *Terebratula*, *Rhynchonella*, and *Spiriferina*. Invariably these faunas are dominantly molluscan. Cephalopods of the families Ophiceratidae, Otoceratidae, and Proptychitidae make up a large proportion of the entire fauna in all regions except the United States. The pelecypods, although

TABLE 2.—*Distribution of the Dinwoody Fauna*

Species	*Lingula* Zone	*Claraia* Zone	Woodside Tongue Equivalent	Ussuri Bay	East Greenland		Alps (Seis)
					Otoceran	Gyronitan	
Discophiceras subkyokticum Spath	X	.....	.....	.....	X	.....	.....
Metophiceras subdemissum Spath	X	.....	.....	.....	X	.....	.....
Mentzelia sp.?	X	.....	.....	.....	.....	.....	.....
Spiriferina mansfieldi Girty	X	.....	.....	?	.....	.....	?
Mytilus? postcarbonicus Girty	X	.....	.....	.....	.....	.....	.....
Pleurophorus? bregeri Girty	X	.....	.....	.....	.....	.....	.....
Lingula borealis Bittner	X	X	.....	X	.....	.....	X
Myalina putiatinensis Kiparisova	X	.....	.....	X	.....	.....	.....
Myalina spathi, n. sp	X	X	X	.....	X	X	.....
Claraia clarai occidentalis, n. var	.....	X	.....	.....	.....	.....	?
Claraia stachei Bittner	X	X	X	.....	X	X	X
Claraia mulleri, n. sp	.....	X	.....	.....	.....	.....	.....
Eumorphotis multiformis Bittner	.....	X	.....	X	X	.....	X
Eumorphotis iwanowi Bittner?	.....	.....	X	X	.....	.....	.....
Myophoria, sp.?	.....	X	X	.....	.....	.....	.....
Anodontophora fassaensis Wissmann	.....	X	.....	X	.....	X	X
Anodontophora canalensis Catullo?	.....	X	.....	X	X	.....	X
Gervilleia ussurica Kiparisova	.....	X	.....	X	.....	.....	.....
Terebratula margaritowi Bittner	.....	.....	X	X	.....	.....	.....
Bellerophon bittneri, n. sp	.....	X	.....	X	.....	.....	.....

numerically important, are generally referred to eight or ten genera, of which the most diagnostic is the pectinoid *Claraia*. Gastropods are represented by only a few poorly known forms, including the last survivors of the Paleozoic Bellerophons.

When the earliest Triassic seas spread across western North America, judging from the character of the Permian-Triassic boundary, they must have transgressed an exceedingly flat surface. If so, this fact, together with the circumstantial evidence afforded by the incredible numbers of *Lingula* shells found in the Dinwoody formation, suggests very shallow waters. The peculiar facies of the early Eo-Triassic faunas may be simply an expression of a uniform and widely distributed environment which simultaneously characterized vast areas in North America, Europe, and Asia. The Dinwoody fauna is about as diverse as the lower Werfen fauna of the Alps and is nearly as varied as the East Greenland faunas from the Lower Triassic, in which fossils other than ammonoids are not well represented. The Dinwoody fauna is strikingly like that from the Ussuri district, except that *Claraia* is apparently unknown in the latter area.

Occurrences of the Dinwoody fauna are summarized in Table 2.

SYSTEMATIC PALEONTOLOGY

BRACHIOPODA

Genus *Lingula*

Lingula borealis Bittner

(Plate 2, figures 1-4)

1899. *Lingula borealis* BITTNER, Com. Géol. Mém., St. Petersburg, vol. 7, p. 25, pl. 4, figs. 1-7, 8?

This form is represented in the collections by hundreds of individuals which exhibit considerable variation in size and some variation in form. Although Bittner sought to distinguish his species from others by the nearly parallel lateral margins there is reason to doubt that this character is invariable. The majority of full-sized individuals, such as the illustrated specimens, commonly taper slightly toward the rear of the shell. These same specimens, as revealed by the growth lines, had subparallel margins in their earlier ontogeny.

OCCURRENCE: Rare in the uppermost beds of the basal siltstone of the Dinwoody formation, very abundant in the *Lingula* zone throughout western Wyoming, southeastern Idaho, and southwestern Montana, and rare in the *Claraia* zone.

Genus *Spiriferina*

Spiriferina mansfieldi Girty?

(Plate 2, figure 9)

1927. *Spiriferina mansfieldi* GIRTY, U. S. Geol. Survey, Prof. Paper 152, p. 436, pl. 30, figs. 17-19.

Represented by a single specimen, an incomplete mold of a ventral valve. There are six plications on each side of the median sulcus, unlike the specimens described

from the Thaynes formation by Girty. However, in absence of adequate material, it is not possible to determine whether or not the number of plications in *S. mansfieldi* is uniformly five or six, as indicated in Girty's description. The writers' specimen is 6 mm. long and 9 mm. wide.

OCCURRENCE: From bed 2, *Lingula* zone, Green River Lakes, Wyoming (section 9).

Genus *Mentzelia*

Mentzelia sp. ?

(Plate 2, figure 8)

1927. *Ambocoelia*, n. sp. GIRTY, U. S. Geol. Survey, Prof. Paper 152, p. 85.

At a number of localities poorly preserved, small, planoconvex brachiopods occur in silty limestones of the Dinwoody formation. Superficially they resemble *Ambocoelia* or *Crurithyris* from the late Paleozoic. At present it seems more appropriate to refer the species tentatively to the Lower Triassic *Mentzelia*.

OCCURRENCE: Rare in the limestones of the *Lingula* zone, western Wyoming (especially numerous in bed 40, McDougal's Pass, near Afton, Wyoming, section 15) and southeastern Idaho.

Genus *Terebratula*

Terebratula margaritowi Bittner

(Plate 2, figures 5a-b)

1899. *Terebratula margaritowi* BITTNER, Com. Géol. Mém., St. Petersburg, vol. 7, p. 27, pl. 4, figs. 10-15.

Poorly preserved terebratuloids are found at various localities in the limestones of the *Claraia* zone in western Wyoming and southeastern Idaho, but specifically identifiable material has not thus far been collected. At a higher horizon, equivalent to the red-beds tongue of the Woodside formation, well-preserved specimens confidently referred to Bittner's Ussuri Bay species were collected in southwestern Montana and southeastern Idaho.

The figured specimen has a length of 13 mm. and a width of 12 mm. The thickness is 6 mm.

OCCURRENCE: Post-*Claraia* beds near Henry, Idaho, bed 105, sec. 18, T.6 S., R. 41E. and about the same stratigraphic position on Trapper Creek, near Melrose, Montana, sec. 22, T. 2S., R. 10W. (section 20).

GASTROPODA

Genus *Bellerophon*

Bellerophon bittneri Newell and Kummel, n. sp.

(Plate 3, figure 12)

1899. *Bellerophon* sp. BITTNER, Com. Géol. Mém., St. Petersburg, vol. 7, p. 28, pl. 4, figs. 26-28.

Shell small, full-sized individuals rarely exceeding 10 mm. across the shell in the plane of coiling; globose, bilaterally symmetrical, completely involute, with a small umbilicus; surface rugose, owing to crowded and irregular varices of growth; slit-band narrow but prominently raised above the general shell surface; aperture extended at lateral margins in a semicircular crest on each side, siphonal notch shallow in front of the slitband.

Measurements (in mm.) of the holotype are as follows:

Width	Diameter	Breadth slitband
9	10	1

The figured specimen is slightly less globular than the material illustrated by Bittner. However, there is some variation in details of form, and some specimens are as inflated as the Siberian specimens.

OCCURRENCE: Bellerophontids were found at Hidden anticline, Red Canyon, Wind River Mountains, 12 miles southeast of Lander, Wyoming, where the *Claraia* beds rest directly on the Phosphoria formation. The holotype, University of Wisconsin No. 21341, was collected at this locality. Poorly preserved Bellerophons were also found in the *Lingula* beds at Cokeville, Wyoming (section 18).

PELECYPODA

Genus *Claraia* Bittner

Claraia clarai var. *occidentalis* Newell and Kummel, n. var.
(Plate 3, figures 1 and 2)

Shell roughly planoconvex, the right valve being flat, or very slightly convex, obliquely ovate, or subrhombic, with a distinct forward obliquity; surface of posterodorsal area smooth, remainder of shell ornamented by 20 to 30 indistinct radial costae, which increase toward the margin by sporadic intercalation of new costae; a few obscure and irregular concentric corrugations cross the radial ornamentation and extend onto the posterodorsal area; byssal notch well defined, but anterior auricle of right valve nearly obsolete; anterior margin of shell extended into a broad lobe.

Measurements (in mm.) of two right valves are as follows:

	Length	Height	Hinge length
1.	28	24	11
2.	27	24	11

From typical *Claraia clarai* Emmrich, principal guide fossil of the Werfen beds of the Alps, this variety differs chiefly in having less pronounced and less regular concentric undulations of the shell.

OCCURRENCE: Rare in the *Claraia* beds of western Wyoming; bed 38, Martin Creek, NW. ¼ sec. 16, T.38N., R.116W. (section 13), and in lower part of *Claraia* zone at Meadow Creek Canyon, sec. 2, T.3N., R.5W. The holotype, University of Wisconsin No. 21343, was collected at the latter locality.

Claraia stachei Bittner
(Plate 3, figures 5-8)

1900. *Claraia stachei* BITTNER, G. R. A., Jahrb., p. 587.
1930. *Claraia stachei*, SPATH, Meddelelser om Grønland, Bd. 83, p. 46, pl. 9, figs. 1a-d; pl. 10, figs. 5a, b.

This is the most abundant and therefore the most characteristic species of *Claraia* found in the early Triassic beds under consideration. There is some variation in form and ornamentation, but ordinarily the shells are nearly circular in plan and are ornamented on both valves by subdued, intercalate costae, which are more numerous on the left valve. Although a left anterior auricle is lacking, that of the right valve is longer and better defined than it is in the majority of species of

Claraia. Shells of this *Claraia* attain a large size for the genus, attaining a length up to 2 inches and more.

Measurements (in mm.) of representative specimens are as follows:

	Left valves			Right valves		
	Length	Height	Hinge length	Length	Height	Hinge length
1.	47	47	25+	19	19.5	9
2.	18	16	5	19	19	9.5
3.	20.5+	22	10.5	26.5	24	12

Occurrence: Rare in shaly equivalents of the basal siltstone near Montpelier, Idaho; rare in the *Lingula* zone at many localities in western Wyoming and southeastern Idaho; abundant in the *Claraia* zone wherever that zone is fossiliferous; rare in the nonred equivalents of the Woodside tongue, near Henry, Idaho.

Claraia mulleri Newell and Kummel, n. sp.

(Plate 3, figures 3 and 4)

Shell suborbicular, posterodorsal area unornamented, extended in a well-defined, subquadrate auricle, anterior auricle lacking on left valve, but a short auricle is present above a well-defined byssal notch on the right valve; surface of both valves covered by intercalate radial costae, numbering up to 40 or 50 in full-sized individuals. Dimensions (in mm.) of the holotype and a paratype are as follows:

Left valve			Right valve		
Length	Height	Hinge length	Length	Height	Hinge length
38.5	37	18±	36	34	20

Occurrence: Rare in the *Claraia* zone in Gros Ventre Canyon, western Wyoming, sec. 6, T.42N., R.114W. (bed 7, section 10). The holotype is 21357 in the University of Wisconsin collections. The species is named in honor of Dr. Siemon Muller.

Genus *Myalina* De Koninck

Myalina spathi Newell and Kummel, n. sp.

(Plate 3, figure 11)

1930. *Myalina* aff. *schamarae*, Spath, Meddelelser om Grønland, Bd. 83, p. 51, pl. 10, fig. 10; pl. 12, figs. 2-3.
1935. *Myalina* aff. *schamarae*, Spath, Meddelelser om Grønland, Bd. 98, p. 70, pl. 20, fig. 12; pl. 21, fig. 15.
1941. *Myalina* aff. *schamarae*, Newell and Kummel, Am. Jour. Sci., vol. 239, p. 208, pl. 1, fig. 2.

Erect, subquadrate, nearly equivalve, small Myalinas, unornamented by concentric rugae or other ornamentation; shell moderately massive. Dimensions (in mm.) of the holotype, a left valve, are: height, 33.5; length, 21.5; hinge length, 15.

This is one of a compact group of Scythian pelecypods commonly assigned to *Myalina* or to *Mytilus*. Other closely related species are *M. platynotus* White, from the Thaynes group, *M. schamarae* Bittner, from the Ussuri Bay region, and *M. putiatinensis* Kiparisova, from the same district. All these species are alike in being unornamented and in having nonprojecting, although angular, beaks. The species listed above have comparatively small shells, rarely exceeding a height of 2 inches. From similar forms the present species is distinguished by the erect shape and by the nearly straight anterior margin.

Occurrence: Widely distributed, and common, in the *Lingula* and *Claraia* zones of the Dinwoody formation in southwestern Montana, southeastern Idaho, and western Wyoming; rare in the beds immediately below the Thaynes, near Henry, Idaho.

The holotype, University of Wisconsin No. 20870, was collected from the *Claraia* zone, near Melrose, Montana, sec. 22, T. 2 S., R. 10W. (bed 9 ? float, section 20).

Myalina putiatinensis Kiparisova

(Plate 3, figures 9 and 10)

1899. *Myalina vetusta,* Bittner, Com. Géol. Mém., St. Petersburg, vol. 7, p. 17, pl. 4, figs. 17-19.
1938. *Myalina putiatinensis,* Kiparisova, Inst. Géol. Acad. Sci. USSR, Travaux, vol. 7, p. 292, pl. 6, figs. 10-12.

This distinctive oblique *Myalina* is confidently identified in the Dinwoody faunas. Measurements (in mm.) of representative specimens are as follows:

Left valve					Right valve				
Length	Height	Hinge	Length	Greatest Dimension	Length	Height	Hinge	Length	Greatest Dimension
21.5	30	16	43.8		20.5	39	15	41.5	

Occurrence: *Lingula* zone, rare at Green River Lakes, Wyoming, SW. ¼ sec. 36, T. 39 N., R. 109 W. (bed 2, section 9); and rare at Gros Ventre Canyon, Wyoming, sec. 6, T. 42 N., R. 114 W. (bed 2, section 10).

Genus *Mytilus*

Mytilus ? postcarbonica (Girty)

1927. *Myalina postcarbonica,* Girty, U. S. Geol. Survey, Prof. Paper 152, p. 442, pl. 30, figs. 34 and 35.

A single specimen of Girty's species, collected at the original locality, occurs in the writers' collections.

Occurrence: *Claraia* zone, Dinwoody formation, Montpelier Canyon, Idaho, W. ½ sec. 31, T. 27 N., R. 45 E. (bed 28, section 19).

Genus *Pleurophorus*

Pleurophorus ? bregeri Girty

(Plate 2, figure 12)

A few fragmentary specimens referred to this species were found at a single horizon and locality, the same that yielded Girty's types. There is some variability within the species, and the figured specimen is shaped differently from the type specimen, as illustrated by Girty. Indeed, the three species of *Pleurophorus* described as new by Girty—*P. bregeri, P. rotundus,* and *P. similis*—came from a single horizon and locality and therefore may well be variants of a single species. An adequate collection would reveal whether or not they completely intergrade.

Occurrence: *Claraia* zone of the Dinwoody formation, Montpelier Canyon, Idaho, W. ½ sec. 31, T. 27 N., R. 45 E. (bed 28, section 19).

Genus *Eumorphotis* Bittner

Eumorphotis multiformis (Bittner)

(Plate 2, figure 10)

1899. *Pseudomonotis multiformis,* Bittner, Com. Géol. Mém., St. Petersburg, vol. 7, p. 10, pl. 2, figs. 15-22.
1935. *Pseudomonotis (Eumorphotis) multiformis,* Spath, Meddelelser om Grønland, Bd. 98, p. 74, pl. 22, fig. 8.
1938. *Pseudomonotis (Eumorphotis) multiformis,* Kiparisova, Inst. Géol. Acad. Sci. URSS, Travaux, vol. 7, p. 224, pl. 2, figs. 4, 9, 12; pl. 3, figs. 2-4.

Several characteristic left valves of this species are in the collections from the *Claraia* zone and higher horizons. There is no close relationship between this species

and the genotype of *Pseudomonotis*, consequently *Eumorphotis* is tentatively elevated to the rank of a genus from its position as a subgenus under *Pseudomonotis*. The relationship to the Paleozoic genus *Limipecten* is much closer. In fact, it is very possible that some of the species commonly listed under *Limipecten* should be referred to *Eumorphotis* or vice versa.

Measurements (in mm.) of characteristic left valves of *E. multiformis* are as follows:

	Height	Length	Hinge Length
1.	42.5	35.5	31
2.	19.0	14.5	12.5

OCCURRENCE: Locally common in the *Claraia* zone, Montpelier Canyon, Idaho (bed 79, section 19), near Henry, Idaho (bed 79, section 21), and near Melrose, Montana (bed 2, section 20).

Eumorphotis iwanowi Bittner ?

1899. *Pseudomonotis iwanowi*, BITTNER, Com. Géol. Mém., St. Petersburg, vol. 7, p. 8, pl. 1, figs. 1-9.

A few very fragmentary left valves collected may belong to this species. They conform rather closely with juvenile stages shown in growth lines of Bittner's figured specimens. The distinguishing features of the species is the early obsolescence of the radial costae, which disappear completely when the shell is half grown. Full-sized Siberian specimens attain a height of more than 2½ inches, whereas the writers' specimens are scarcely more than an inch in height. Larger and more mature specimens must be obtained before the American material can confidently be assigned to this species.

OCCURRENCE: Rare in the uppermost pre-Thaynes beds near Henry, Idaho (bed 79, section 21).

Genus *Anodontophora* Cossmann

Anodontophora fassaensis (Wissmann)

(Plate 2, figure 13)

1841. *Myacites fassaensis*, WISSMANN, in Graf zu Münster, Beitr. Geogn. sudostl. Tirols, IV, p. 9, Taf. 16, fig. 2.
1899. *Anodontophora fassaensis*, BITTNER, Com. Géol. Mém., St. Petersburg, vol. 7, p. 22, pl. 3, figs. 28-33.
1930. *Anodontophora* cf. *fassaensis*, SPATH, Meddelelser om Grøland, Bd. 83, p. 55, p. 11, fig. 2.
1938. *Anodontophora fassaensis*, KIPARISOVA, Inst. Géol. Acad. Sci. URSS, Travaux, vol. 7, p. 267.

Numerous specimens from the *Claraia* zone of the Dinwoody formation fall within the rather broad limits generally ascribed to this species. The writers' specimens are rough molds, so that surface details generally are not well shown. Furthermore, the majority of specimens are somewhat crushed or otherwise deformed. However, a number of specimens are moderately well preserved. Dimensions (in mm.) of characteristic Dinwoody specimens are as follows:

	Length	Height
1.	16	10.5
2.	11	7
3.	13	10
4.	15	10
5.	12	8±

OCCURRENCE: Common in the *Claraia* zone of the Dinwoody formation in western Wyoming and southeastern Idaho.

Anodontophora canalensis (Catullo)?

1848. *Tellina canalensis*, CATULLO, Mem. geogn. pal. Alpi Venete, p. 56, pl. 4, fig. 4.
1899. *Anodontophora canalensis*, BITTNER, Com. Géol. Mém., St. Petersburg, vol. 7, p. 23, pl. 3, figs. 34-38.
1931. *Anodontophora canalensis*, SPATH, Meddelelser om Grønland, Bd. 83, p. 54, pl. XII, fig. 10, 11, 13.
1938. *Anodontophora canalensis*, KIPARISOVA, Inst. Géol. Acad. Sci. URSS, Travaux, vol. 7, p. 269.

Fragmentary and otherwise poorly preserved specimens from the Dinwoody formation probably belong to this well-known pelecypod species. Further collecting may yield more satisfactory material.

OCCURRENCE: Questionably identified from the *Claraia* zone at a few localities in western Wyoming.

Genus *Gervilleia* Defrance

Gervilleia ussurica Kiparisova

1899. *Gervilleia* cf. *exporrecta*, BITTNER, Com. Géol. Mém., St. Petersburg, vol. 7, p. 15, pl. 3, figs. 1-5.
1930. *Gervilleia* aff. *exporrecta*, SPATH, Meddelelser om Grønland, Bd. 83, p. 48, pl. 10, figs. 11a-b; pl. 12, fig. 4.
1935. *Gervilleia* aff. *exporrecta*, SPATH, Meddelelser om Grønland, Bd. 98, p. 69, pl. 22, figs. 9a-c.
1938. *Gervilleia ussurica*, KIPARISOVA, Inst. Geol. Acad. Sci. URSS, Travaux, vol. 7, p. 241, pl. 6, figs. 5-8.

A small pterioid pelecypod like Kiparisova's species is locally common in the upper beds of the Dinwoody formation in the western mountain ranges of Wyoming and in southeastern Idaho. The specimens are generally poorly preserved, so that critical generic characters of the hinge are not discernible. Considering the associated fauna it is highly probable that the writers' species is the same as that described from Siberia by Kiparisova and from Greenland by Spath.

OCCURRENCE: Common in the *Claraia* zone of the Dinwoody formation in western Wyoming and southeastern Idaho.

CEPHALOPODA

Genus *Ophiceras*

Subgenus *Discophiceras* Spath

Ophiceras (Discophiceras) subkyokticum Spath

(Plate 2, figure 6)

1935. *Ophiceras (Discophiceras) subkyokticum*, SPATH, Meddelelser om Grønland, Bd. 98, p. 34, pl. 1, figs. 3a-b; pl. 8, figs. 6a-b; pl. 12, figs. 6a-b; pl. 13, figs. 6a-b, 9; pl. 18, figs. 2a-b.
1941 *Ophiceras (Discophiceras) subkyoticum*, NEWELL AND KUMMEL, Am. Jour. Sci., vol. 239, p. 207, pl. 1, figs. 4a-b.

A single fragmentary specimen, referred to Spath's species, consists of approximately one-third of the outer volution minus the living chamber. The left side of the specimen is partly destroyed by weathering, but enough remains for a satisfactory restoration of the whorl section. At the adapertural end the specimen

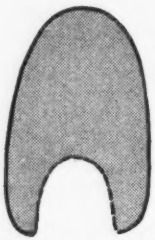

FIGURE 3.—*Whorl section of* Discophiceras subkyokticum *Spath*
Partially restored; from near the adapertural end of the conch.

measures 19 mm. in height, as measured from the umbilical seam to the venter. The greatest transverse dimension, measured through the outer rim of the umbilicus, is estimated at 36.6 mm.

Anyone familiar with early Triassic ophiceratoids is likely to be impressed by the similarity of different species and genera. A thoroughly satisfactory identification should be based on several complete specimens, in order to define the inherent

FIGURE 4.—*Suture pattern of* Discophiceras subkyokticum *Spath*
This pattern is a composite drawing from an intermediate part of the figured specimen.

variability. However, with due consideration to the fragmentary nature of the present specimen, it agrees in all essential respects with Spath's species.

OCCURRENCE: Represented by one specimen collected from the *Lingula* zone 150 feet above the base of the Dinwoody (bed 1B.5, section 20), near the west line of sec. 22, T. 2S., R.10W., about 9 miles west of Melrose, Montana.

Subgenus *Metophiceras* Spath
Ophiceras (Metophiceras) subdemissum Spath
(Plate 2, figure 7)

1935. *Ophiceras (Metophiceras) subdemissum* SPATH, Meddelelser om Grønland, bd. 98, p. 35, pl. 13, fig. 3; pl. 17, figs. 3a-b; pl. 18, figs. 3a-b; pl. 19, fig. 2.

One poorly preserved specimen referred to this species was found in the Dinwoody formation at Montpelier, Idaho. The specimen has been badly weathered and crushed but does show some of the essential characters of Spath's species. The

sutures are too poorly preserved to show denticulations on the lobes. The specimen measures 27 mm. through its greatest transverse diameter; the height of the outer whorl from the umbilical seam to the venter is 7 mm.

FIGURE 5.—*Suture pattern of* Metophiceras subdemissum *Spath*

The suture of the species figured by Spath has faint denticulations on the lobes; these are not present on the writers' specimen possibly because of poor preservation.

The problem of identification here is similar to that of *Discophiceras subkyokticum* —poor preservation and lack of sufficient material prevent positive identification.

OCCURRENCE: The single specimen in the writers' collection came from the *Lingula* zone, Dinwoody formation, Montpelier Canyon, Idaho (bed 12, section 19).

STRATIGRAPHIC SECTIONS

4.—BULL LAKE CANYON, WIND RIVER MOUNTAINS, WYOMING
NE¼ sec. 32, T. 3N., R. 3W.

Bed No.		Feet
	Red Peak formation (section incomplete)	
	Sandstone, red, cliff-former	
69	Shale, dark brick-red, gray beds, intermingled, largely covered, contains several 1-inch gypsums	22.0
68	Siltstones, gray, shaly, fairly resistant	0.4
67	Shale, dark brick-red, silty, few thin inconspicuous green shales present	14.0
66	Shale, red and gray, hard, silty	1.0
65	Siltstone, light greenish-gray, calcareous, some red shale present	0.5
64	Siltstone, shaly, brick-red	1.0
63	Siltstone, shaly, hard, makes good bench	2.0
62	Shale, dark brick-red, silty	6.0
61	Siltstone, fine-grained, brick-red, irregular and platy, makes obscure bench	1.0
60	Shale, dark brick-red, flaky, silty, partly covered	22.0
59	Siltstone, coarse-grained, light red	8.5
	Underlying rocks are silty; from here up they include some coarse siltstone or fine sandstone; gradational contact	
58	Shale, dark brick-red, intermingled with irregular streaks of blue-gray shale	1.0
57	Shale, dark brick-red, lithologically similar to underlying shales	19.0
56	Shale, olive, flaky, characterized by closely spaced joints	12.0
55	Shale, gray, flaky, contains several thin gypsums	11.0
54	Shale, as preceding, but less gypsum	8.5
53	Satin spar	0.1
52	Shale, as No. 55	4.0
51	Gypsums, thin, and olive shales	0.6

Bed No.		Feet
50	Same as No. 47	5.5
49	Satin spar	0.2
48	Same as No. 47	5.5
47	Shale, olive, flaky, broken by thin joints	5.5
46	Shale, same as 45, with 1-inch satin spar at top	4.0
45	Shale, olive to buff, soft, flaky, contains 2-inch bed satin spar at top	3.0
		158.3
	Dinwoody formation	
44	Siltstone, brown, drab, calcareous, with interbedded gypsums, forms top of bench forming unit	2.0
43	Siltstone, sandy, gray	2.5
42	Shale, dark-gray	2.5
41	Siltstone, buff, silty, pitted weathered surface	1.4
40	Shale, buff, silty	1.5
39	Siltstone, massive, gray	5.5
38	Siltstone, gray to buff, shaly	6.0
37	Siltstone, gray, massive, some parts shaly	3.5
36	Shale, gray, silty	0.8
35	Siltstone, gray, thin-bedded	0.6
34	Siltstone, brown, sandy, massive, part shaly, contains small pelecypods	8.5
33	Shale, brown, sandy toward top	8.0
32	Siltstone, gray, shaly, weathers buff	1.0
31	Shale, gray, silty	2.5
30	Siltstone, gray, massive	0.3
29	Shale, gray, silty	5.0
28	Shale, gray, silty, contains 2-inch band of siltstone in center	0.8
27	Siltstone, gray, massive, contains iron concretions	0.5
26	Shale, gray, silty, thin-bedded, contains pelecypods	5.0
25	Limestone, gray, silty, contains small pelecypods	2.0
24	Shale, gray, silty, two siltstone bands 2 inches thick near middle	3.8
23	Limestone, same as 19	0.9
22	Shale, same as 20	0.3
21	Limestone, same as 19, contains *Lingula*	0.7
20	Shale, gray, silty	0.3
19	Limestone, gray, silty, irregular, massive, contains *Lingula*	1.7
18	Shale, gray, silty, contains few thin siltstones	2.8
17	Limestone, gray, massive, with poorly preserved fossils, thin siltstone at bottom	0.9
16	Shale, gray, same as 14	2.0
15	Limestone, gray, massive, contains poorly preserved pelecypods, *Lingula*	1.4
14	Shale, gray, silty	0.8
13	Limestone, gray, massive, bottom silty, contains *Lingula*	2.4
12	Shale, dark-gray, silty	0.3
11	Siltstone, gray, thin-bedded, weathers brown	0.5
10	Shale, buff, silty	1.5
9	Siltstone, gray, sandy, massive, beds 3 to 4 inches thick	2.5
8	Shale, green and gray, silty	1.0
7	Siltstone, buff, massive	3.5
6	Siltstone, buff, massive	1.0
5	Shale, buff, silty	0.3
4	Siltstone, buff, sandy, massive, with shale partings	0.2
3	Shale, buff, silty	0.1
2	Shale, greenish	0.2
1	Clay, grayish-buff, gypsum or calcite crystals present, sandy	0.5
	Total Dinwoody	89.5
	Phosphoria formation	

5.—Dinwoody Canyon, Wind River Mountains, Wyoming
S.½ sec. 31, T. 5N., R. 5W.

Bed No.		Feet
	Red Peak formation (section incomplete)	
22	Sandstone, red, fine-grained, with red shale partings, forms scarp	60.0±
21	Sandstone, red, platy, and red soft shales, a few thin olive partings	35.0
20	Shale, olive, platy, and light olive, fine-grained sandstone or siltstone	15.0
19	Sandstone, olive, more or less massive, soft	6.0
18	Shale, olive, platy, several thin (less than ½ inch thick) gypsums	15.0
17	Sandstone, light olive, more or less massive, fine-grained, soft	4.0
16	Shale, olive, soft, flaky, silty, thin gypsum beds (less than ½ inch thick) every 5 feet or so	20.0
15	Like 14 only better exposed	25.0
14	Shale, olive, soft, not platy, exposure only fair	25.0
		205.0
	Dinwoody formation	
13	Sandstone, gray, fine-grained, platy, weathers brown, irregularly rippled, calcareous at top	25.0
12	Sandstone, gray, fine-grained, shaly, weathers brown, contains pelecypods at several horizons	7.0
11	Sandstone, gray to brown, shaly, contains pelecypods and calcite geodes at top	4.0
10	Sandstone, light-gray, fine-grained, weathers brown, contains pelecypods at base, shaly above	9.0
9	Shale, gray, platy, more silty at base	10.0
8	Shale, silty, contains calcite geodes	0.4
7	Siltstone, light gray, fine-grained, shaly	5.0
6	Shale, olive and thin olive siltstones	20.0
5	Limestone, brown, gray, where fresh, contains *Lingula*	1.0
4	Sandstone and shale, gray and buff, calcareous	3.5
3	Limestone, gray, granular, weathers brown, contains poor pelecypods, *Lingula*	2.0
2	Sandstone, red, shaly, fine-grained, contains *Lingula*	2.0
1	Sandstone, bituminous, cross-bedded	1.0
	Total Dinwoody (as redefined)	89.9
	Phosphoria formation	
	Limestone, cherty, weathered zone of tripoli at top.	

6.—Little Red Canyon, Wind River Mountains, Wyoming
T.40N., R.105W.

Bed No.		Feet
	Red Peak formation (section incomplete)	
28	Shale, greenish-gray, partly covered	117.0
	Dinwoody formation	
27	Siltstone, like 26, massive, gets more sandy toward top, weathers shaly	19.0
26	Siltstone, dark-gray, hard, thin-bedded	2.0
25	Shale, dark bluish-gray	0.7
24	Siltstone, gray, thin-bedded, some parts shaly, weathers dark brown	2.5
23	Shale, dark bluish-gray	3.0
22	Siltstone, Sandy, massive, weathers dark brown and shaly, contains small pelecypods	5.5
21	Shale, dark bluish-gray	0.6
20	Siltstone, gray, sandy, weathers dark brown, upper 2½ feet calcareous, a coquina of small pelecypods	4.5
19	Siltstone, gray, shaly, weathers dark brown, contains small pelecypods	1.0
18	Shale, buff, sandy	0.6
17	Limestone, gray, massive, silty, contains small pelecypods	0.7

Bed No.		Feet
16	Siltstone, gray, sandy, thin-bedded, weathers shaly, contains small pelecypods	3.0
15	Siltstone, gray, massive, hard, contains small pelecypods	0.9
14	Siltstone, gray, shaly, weathers buff, contains small pelecypods	2.4
13	Siltstone, gray to brown, massive, weathers dark brown or buff, contains small pelecypods	3.3
12	Shale, gray, silty, contains small pelecypods	1.0
11	Siltstone, gray, massive, lower part weathers shaly	1.0
10	Shale, gray, silty, contains small pelecypods	1.3
9	Siltstone, gray, weathers buff, massive, contains small pelecypods	0.7
8	Shale, bluish-gray	4.0
7	Shale, gray, silty, weathers buff	2.5
6	Covered	5.5
5	Limestone, light-gray, shaly in parts, weathers dark brown, contains *Lingula* and *Myalina*	6.0
4	Siltstone, gray, thin-bedded, shaly, weathers buff, sandy in parts, contains *Lingula*	6.5
3	Shale, gray, silty, weathers buff	1.0
2	Shale, bluish-gray	0.5
1	Shale, sandy, soft, iron-stained bed at base, remainder gray	0.4
	Total Dinwoody	80.1

Phosphoria formation

Limestone, shaly.

7.—Red Canyon, Owl Creek Mountains, Wyoming

SW¼ sec. 24, T.7N., R.4W.

Bed No.		Feet
	Red Peak formation (section incomplete)	
42	Sandstone, brick-red, fine-grained, makes heavy bench, slabby beds, separated by shale partings	...
41	Shale, dark brick-red, silty, blocky, contains a few thin gray beds, partially covered	60.0
40	Siltstone, gray, hard, platy	1.0
39	Shale, dark brick-red, silty	3.5
38	Shale, gray, hard, calcareous, silty, contains few red streaks	1.5
37	Shale, dark brick-red, silty	31.5
36	Shale, olive-gray, silty	1.0
35	Shale, dark brick-red, silty, contains a few thin gray streaks, sharp contact with underlying shale	117.0
34	Shale, olive, silty, flaky	8.0
33	Shale, gray, hard, slaty, looks like varves, contains small crystals of pyrite	2.5
32	Shale, olive, silty, poorly bedded, contains few gypsum beds	16.0
		242.0
	Dinwoody formation	
31	Siltstone, gray, hard, platy, rippled	3.0
30	Shale, gray, argillaceous to silty	2.0
29	Siltstone, bluish-gray, platy, rippled	4.0
28	Limestone, light gray, fine-grained, silty, filled with small pits	3.0
27	Shale, bluish to buff, hard, platy, silty, calcareous, rippled	4.0
26	Siltstone, bluish-gray, hard, calcareous, contains several coquina bands of small pelecypods up to 6 inches thick	7.5
25	Siltstone, gray, calcareous, hard, in beds up to 1 foot thick	3.0
24	Siltstone, gray, calcareous, hard, cross-bedded, contains band of calcite geodes up to 2 inches in diameter at top	1.0
23	Limestone, brown to gray, silty, a coquina of small pelecypods	0.7

Bed No.		Feet
22	Siltstone, gray, calcareous, hard, contains no fossils	1.0
21	Limestone, brown, silty, a coquina of small pelecypods	1.0
20	Siltstone, gray, calcareous, hard, unfossiliferous	3.5
19	Shale, green, silty	0.8
18	Limestone, buff to brown, silty and argillaceous, very hard	1.5
17	Siltstone, calcareous and argillaceous gray shale	4.0
16	Limestone, buff to blue, argillaceous, silty, very hard	0.7
15	Shale, gray, argillaceous, contains some thin siltstones	2.5
14	Limestone, blue, argillaceous, in beds up to 6 inches thick, separated by shale partings	12.0
13	Limestone, bluish-gray, blocky, silty, contains abundant pelecypods, blue shale at top	4.0
12	Shale, gray to buff, hard, calcareous, contains few pelecypods	5.0
11	Limestone, gray, brittle, filled with small pelecypods	1.0
10	Limestone, brown to buff, blocky, silty, contains much shale and argillaceous siltstone, contains *Lingula* and pelecypods	7.0
9	Shale, olive, argillaceous, contains some thin siltstones	1.5
8	Limestone, blocky, hard, silty, weathers brown	2.0
7	Shale, olive, silty, contains several hard thin siltstones	1.0
6	Siltstone, brown to buff, hard, "storm roller" structure, flat top, uneven base, contorted beds	3.0
5	Shale, olive, silty, contains thin siltstones	0.9
4	Siltstone, brown, hard, calcareous, "storm roller" structure, flat top, irregular and contorted base, contains small pelecypods	1.0
3	Siltstone, shaly, with a few sandy beds, contains small pyrite concretions that weather to limonite	5.0
2	Sandstone, buff to orange, calcareous, blocky, rippled, fine-grained, in hard beds up to 8 inches thick	3.0
1	Shale, buff, silty, contains thin irregular siltstones, no fossils observed	21.0
	Total Dinwoody	110.6

Phosphoria formation

Limestone, white, cherty, unfossiliferous, relatively thick.

8.—Blue Holes Canyon, Wind River Mountains, Wyoming
T.40N., R.105W.

Bed No.		Feet
	Red Peak formation (section incomplete)	
43	Sandstone, red, fine-grained, massive, mottled gray at base, basal bed of cliff-forming sandstone unit	...
42	Shale, dark brick-red, silty, poorly exposed, contains few siltstone beds in upper part	75.0
41	Shale, gray, silty	45.0
		120.0
	Dinwoody formation	
40	Siltstone, olive-gray, thin-bedded, hard	1.5
39	Siltstone, shaly, hard	5.0
38	Siltstone, gray to buff, thin-bedded	1.0
37	Shale, olive-gray, silty	14.5
36	Siltstone, olive-gray, thin-bedded, inconspicuous ledge maker	1.5
35	Shale, olive, silty	18.0
34	Siltstone, buff-gray, thin-bedded, weathers dark brown	9.5
33	Siltstone, buff and gray, sandy and massive in lower part, thin-bedded in upper part	3.5
32	Shale, buff, silty	0.3
31	Siltstone, gray, sandy, massive, weathers dark brown	3.5

Bed No.		Feet
30	Shale, gray	1.2
29	Siltstone, gray, massive, weathers dark brown	3.0
28	Shale, greenish-gray, some gypsum present	1.0
27	Siltstone, gray-buff, weathers dark brown, contains small pelecypods	3.0
26	Shale, greenish-gray, contains a few thin siltstone beds	2.0
25	Siltstone, gray, orange in certain beds, weathers dark brown, contains small pelecypods	3.8
24	Shale, gray, silty, weathers buff	1.0
23	Siltstone, gray, massive, weathers dark brown, contains small pelecypods	2.7
22	Siltstone, orange-buff, upper part gray, massive, weathers shaly and dark brown	2.5
21	Siltstone, gray, massive, weathers dark brown, contains small pelecypods	1.3
20	Shale, gray	0.4
19	Siltstone, orange-buff, sandy, weathers dark brown, thin-bedded, contains small pelecypods	1.5
18	Shale, gray, silty	0.3
17	Siltstone, orange-buff, weathers dark brown, shows spheroidal weathering, contains small pelecypods	2.5
16	Shale, buff, silty	0.4
15	Siltstone, gray, weathers shaly and spherically, contains small pelecypods	3.6
14	Shale, green	1.7
13	Siltstone, gray, massive, weathers shaly	0.7
12	Shale, green	0.7
11	Siltstone, grayish-buff, weathers shaly, thin gypsum beds present	1.0
10	Siltstone, contains orange and gray laminations, fine-grained, weathers shaly	0.5
9	Siltstone, gray, weathers shaly	2.0
8	Covered	5.0
7	Siltstone, orange-buff, shaly, contains *Lingula* and small pelecypods	3.0
6	Limestone, grayish-white, silty, massive, weathers dark brown, contains *Lingula* and small pelecypods	5.0
5	Shale, buff, silty	2.0
4	Siltstone, sandy, thin-bedded, weathers dark brown	5.0
3	Siltstone, buff, shaly, weathers dark brown	2.5
2	Shale, gray	1.5
1	Shale (?) covered	2.0
	Total Dinwoody	120.8

Phosphoria formation

Limestone.

9.—Green River Lakes, Wind River Mountains, Wyoming

West Side of Sheep Mountain

SW. ¼ sec. 36, T. 39N., 109W.

Bed No.		Feet
	Red Peak formation (section incomplete)	
20	Sandstone, red, heavy, slabby, fine-grained, makes bench	....
19	Shale, dark brick-red, silty, partly covered	100.0
18	Shale, gray, silty	0.5
17	Shale, dark brick-red, silty	20.0
16	Shale, red, partly covered	60.0
15	Shale, contains few siltstone bands, gray, upper boundary told by color change in soil	30.0
		210.5

Bed No.		Feet
	Dinwoody formation	
14	Siltstone, slabby, calcareous, brittle, weathers brown, fossiliferous, upper part shaly	20.0
13	Shale, partly covered	5.0
12	Siltstone, dark-buff, massive, exhibits spheroidal weathering, weathers dark brown, fossiliferous, shaly in parts, more thin-bedded, shaly and limy toward top	30.0
11	Siltstone, gray, shaly, alternating with silty shales, *Claraia* zone 5 feet above base	24.0
10	Shale, gray, silty	2.0
9	Siltstone, gray, thin-bedded	1.0
8	Shale, olive-gray, silty	2.0
7	Siltstone, gray, beds 4 to 6 inches thick, weathers brown	14.0
6	Siltstone, slabby, with calcite geodes	1.0
5	Shale, partly covered	4.0
4	Siltstone, rippled, calcareous	0.5
3	Shale, silty, partly covered	9.0
2	Limestone, gray, sandy, fossiliferous, *Lingula* bed near top	2.0
1	Sandstone, fine-grained, calcareous, basal unit has flat hard clay and calcitic shaly pebbles	4.0
	Total Dinwoody	118.5
	Phosphoria formation	
	Limestone, thin-bedded, platy, upper layer sugary, has streaks of drusy quartz	

10.—Gros Ventre Canyon, Wyoming
1 mile NW. of Gros Ventre Slide
NW. ¼ SE. ¼ sec. 10, T. 42N., R. 114W.

Bed No.		Feet
	Sundance formation (section incomplete)	
	Nugget formation (?)	
107	Sandstone, red, cliff forming, estimated at	65.0
	Thaynes formation (Ankareh formation absent?)	
106	Covered	75.0
105	Sandstone, white to pink, weathers to fretted profile	3.0
104	Covered	11.0
103	Sandstone, red, calcareous	8.0
102	Covered, probably shale	62.0
101	Limestone, maroon, sandy	1.0
100	Covered, red, sandy shale	16.0
99	Limestone, pink to maroon, shaly, cavernous, nodular	1.5
98	Limestone, white, laminated, streaked red, contains small quartz geodes	7.0
97	Shale, red, silty, limy	11.0
96	Limestone, light-gray, shaly, very platy	7.0
95	Shale, light-gray, calcareous	3.5
94	Limestone, light-gray to white, shaly	1.0
93	Shale, red and green	1.5
92	Limestone, shaly, yellowish	1.5
91	Limestone (basal bed of Alcova?), gray, silty, very thin-bedded, lower part more silty	1.5
90	Shale, maroon in lower part, upper part olive	8.0
89	Sandstone, gray, soft	1.0
88	Covered	31.0
87	Sandstone, shaly	5.0
86	Sandstone, red	3.0
85	Shale, red	5.0
84	Sandstone, pink, massive, calcareous	10.0
83	Covered, shale and red shaly siltstone	35.0

Bed No.		Feet
82	Sandstone, brown, massive, limy	1.0
81	Siltstone, red, shaly	48.0
80	Sandstone, pink, limy, very massive, weathers rounded	4.0
79	Sandstone, red, shaly	15.0
78	Limestone, maroon and gray, conglomeratic, sandy	5.0
77	Limestone, gray, crystalline, very irregular and conglomeratic, contains sandstone pebbles	2.5
76	Limestone, red, sandy and red shale	7.0
75	Limestone, light-gray, granular, interbedded gray shale	8.0
	Total Thaynes	400.0
	Woodside formation	
74	Shale, red, sandy	7.0
73	Shale and sandstone, blue-gray	1.0
72	Sandstone, red, shaly	4.0
71	Shale, blue-gray and red, sandy	2.0
70	Sandstone, red, shaly	17.0
69	Sandstone, gray, blocky	2.0
68	Sandstone, red, massive, cliff-forming, rippled, even-bedded	29.0
67	Sandstone, red, slabby, makes slope	16.0
66	Shale and shaly sandstone, covered	56.0
65	Sandstone, gray, slabby, fine-grained	0.6
64	Shale and slabby sandstone, covered	11.0
63	Sandstone, red, makes bench	1.5
62	Shale, red, sandy	12.0
61	Sandstone, red, shaly	14.0
60	Shale, red, silty, poorly exposed	11.0
59	Sandstone, shaly	8.0
58	Sandstone, red, shaly, medium-grained	20.0
57	Shale, limy, turquoise-colored	0.2
56	Sandstone, red, shaly	8.0
55	Sandstone, red, massive	22.0
54	Shale, red, silty, with green streaks	17.0
53	Sandstone, light-gray to greenish-gray	2.0
52	Sandstone, red, fine-grained	3.5
51	Sandstone, light-gray, slabby	2.0
50	Sandstone, red, with one or two brown streaks	10.0
49	Sandstone, red, massive, upper part light gray	4.5
48	Sandstone, light-gray	2.5
47	Shale, red, and shaly sandstone	6.0
46	Sandstone, red, silty	5.0
45	Shale, red, silty with a few red sandstone bands	20.0
44	Sandstone, red, slabby and red shales	17.0
43	Sandstone, red, fine, massive, makes cliff	12.0
42	Sandstone, red, limy and fine red shales	44.0
41	Sandstone, red, fine and sandy red shales	45.0
40	Shale, gray and red, silty	2.0
39	Shale, dark brick-red, silty	54.0
38	Siltstone, gray, conglomeratic	2.0
37	Shale, maroon	2.0
36	Siltstone, olive	5.0
35	Limestone, olive-gray, fossiliferous	0.5
34	Siltstone and shale, olive-gray	5.0
33	Shale, maroon, silty	4.0
32	Siltstone, calcareous	1.5
31	Shale, dark brick-red, silty	2.0
30	Shale, olive-gray, silty	3.0
29	Siltstone, olive-gray	3.0
28	Shale, silty and siltstone, gray to olive	2.0

Bed No.		Feet
27	Siltstone, calcareous, gray, with poorly preserved fossils	2.0
26	Siltstone, gray, poorly bedded with solution pits	4.0
25	Shale, dark brick-red, silty	20.0
24	Siltstone, red and gray, hard	1.0
23	Shale, dark brick-red, silty, an inconspicuous gray bed in middle	7.0
	Total Woodside	552.8
	Dinwoody formation	
22	Siltstone, gray, platy, fissile	4.0
21	Sandstone, buff to brown, friable, slabby, fine, rippled	8.0
20	Shale, bluish-gray, papery, silty	3.0
19	Limestone, light-gray, fine-grained, has small snails, abundant *Gervilleia, Lingula*	3.0 to 5.0
18	Shale, olive-gray	15.0
17	Siltstone and shale, olive-gray, flaggy	3.5
16	Sandstone, fine, buff, friable	5.0
15	Shale, gray, silty	3.0
14	Limestone, light-gray, nodular, fossiliferous, *Gervilleia*	1.5
13	Shale, gray, silty, upper part light-gray, fissile, hard	9.0
12	Siltstone, buff, soft, a few fossils	17.0
11	Limestone, light-gray, coquinoid, contains *Claraia,* blocky, weathers brown	2.5
10	Siltstone, calcareous, weathers platy and brown, has small calcite vugs	18.0
9	Siltstone, buff-gray, fossiliferous, two zones of "storm rollers", large *Myalina*	3.0
8	Siltstone, gray-buff, platy, shaly, "storm rollers" at base, small calcite vugs	15.0
7	Siltstone, buff, shaly, poorly shown, sparsely fossiliferous	34.0
6	Shale, silty, gray, mainly covered	53.0
5	Limestone, gray, with orange spots, lower part more silty, fossiliferous	3.0
4	Shale, gray, silty, poorly exposed	25.0
3	Limestone, fine-grained with orange spots, blocky, weathers brown, lower part covered, contains *Lingula,* pelecypods	2.5
2	Sandstone, fine-grained, buff, has poorly preserved *Lingula* and pebbles of chert, some clay pebbles	3.0
1	Covered, shale, buff	4.0
	Total Dinwoody	235.0 to 237.0
	Phosphoria formation	
	Limestone, black, cherty, platy, with bryozoans	

11.—Bear Gulch, Hoback Canyon, Hoback Mountains, Wyoming
W. center sec. 25, T. 39 N., R. 115 W.

Bed No.		Feet
	Thaynes formation (section incomplete)	
43	Shale, red, silty, measured up to heavy limestone and siltstone, on mountain top, thickness estimated	150.0 to 200.0
42	Limestone, red, conglomeratic with gray limestone pebbles, subangular and rounded, up to 1 inch in diameter	3.0
41	Limestone, massive, light-gray and pink, weathers white, aphanitic, laminated	12.0
		165.0 to 215.0
	Woodside formation	
40	Covered, red beds	125.0
39	Limestone, bench made by blocks of pink fine limestone, up to 2 feet thick	2.0

Bed No.		Feet
38	Covered, red beds	28.0
37	Limestone, pink, fine-grained, granular, looks as though it is in place, but this is not sure	2.0
36	Covered, red beds	420.0
35	Shale, dark brick-red, silty	30.0
34	Siltstone, olive to gray, calcareous, shaly, makes bench	7.0
33	Shale, dark brick-red, silty	5.0
32	Limestone, gray, dense, platy, contains no fossils, brecciated or conglomeratic	9.0
31	Shale, dark brick-red, silty	2.5
30	Limestone, olive and red, silty, contains no fossils, platy and flaggy	7.0
29	Shale, maroon or dark brick-red	5.5
	Total Woodside	643.0
	Dinwoody formation	
28	Limestone or dolomite, cavernous, punky and chalky	2.0
27	Siltstone and shale, olive	4.0
26	Limestone, light-gray to white, massive, contains no fossils	20.0
25	Limestone, light-gray, silty, massive, cross-bedded, contains small pelecypods and gastropods	30.0
24	Siltstone, gray, shaly	2.5
23	Siltstone, olive-gray to buff, blocky, quartzitic, contains few poorly preserved pelecypods	135.5
22	Shale, olive, silty	2.5
21	Limestone, light-gray to buff, coquinoid, contains *Lingula*	4.0
20	Shale, olive-green, silty	8.0
19	Shale, olive, silty and calcareous, siltstone	4.0
18	Limestone, silty, weathers brown, platy, makes ledge	5.0
17	Siltstone, gray, platy, shaly, calcareous, makes slope	4.5
16	Limestone, bluish-gray, dense, hard, weathers buff to gray, contains good *Lingula*	4.0
15	Shale, gray, silty, unfossiliferous	5.0
14	Limestone, bluish-gray, brittle, weathers whitish-brown, coquina of *Lingula*	11.0
13	Shale, greenish-olive-gray, exhibits concoidal fracture, harder in spots, *Lingula* present	15.0
12	Limestone, grayish-blue, hard, brittle, beds 3 to 4 inches thick, contains good *Lingula*	4.0
11	Siltstone, shaly and silty shales, olive-gray	10.0
10	Limestone, bluish-gray, hard, brittle, contains *Lingula*	0.4
9	Siltstone, shaly, gray, contains *Lingula*	1.5
8	Limestone, grayish-blue, massive, contains *Lingula*	1.0
7	Shale, gray to buff, silty, contains 2 inch limestone band near top with *Lingula*	12.0
6	Limestone, bluish-gray, hard, brittle, contains *Myalina, Claraia, Lingula*	1.0
5	Shale, gray, silty and thin siltstone bands	10.0
4	Limestone, same as 2	0.4
3	Shale, gray, silty	8.0
2	Limestone, bluish-gray, hard, brittle, weathers brown, contains *Myalina, Lingula,* and small pelecypods	2.0
1	Shale, gray, silty	15.0
	Total Dinwoody	322.3
	Phosphoria formation	
	Limestone, bluish-gray, silty, brittle, hard, fine-grained, fetid odor on pounding, contains *Avonia, Composita* 25 feet above base, lower 10 feet forms slope, upper part cliff former	100.0
	Phosphorite, black, oölitic	

12.—Phillips Pass, Teton Range, Wyoming

E. ½ sec. 6, T. 41 N., R. 117 W.

Bed No.		Feet
	Woodside formation (section incomplete)	
	Shale, red, soft, weathered slope	
	Dinwoody formation	
27	Sandstone, olive-colored, slabby	6.0
26	Limestone, oatmeal texture, contains poorly preserved fossils	1.5
25	Covered interval	20.0
24	Sandstone, gray, poorly exposed, fine-grained, weathers brown	10.0
23	Covered	15.0
22	Limestone, bluish-gray, fine-grained, hard, contains fragments of fossils	3.0
21	Siltstone, thin-bedded, forms slope, slight change in fauna, fossils more numerous and varied	25.0
20	Siltstone, brown, massively bedded, contains *Claraia* and *Myalina*	20.0
19	Siltstone, olive, platy, shaly	2.0
18	Same as 17, contains abundant pelecypods	4.0
17	Siltstone, calcareous, massive, quartzitic, "storm rollers" at base	4.0
16	Siltstone, calcareous, blocky, subquartzitic	3.0
15	Shales and siltstones, calcareous, interbedded with limestones which contain *Lingula*	13.0
14	Limestone, contains *Lingula*	4.0
13	Shale, calcareous and silty	2.0
12	Limestone, shaly, weathers brown	5.0
11	Shale, silty, partially covered	3.0
10	Limestone, bluish-gray, fine-grained, hard, contains *Lingula* and shale breaks	9.0
9	Shale, silty, partially covered	82.0
8	Siltstone, gray to buff, slabby	8.0
7	Limestone, gray, fine-grained, contains *Lingula*	4.0
6	Siltstone, buff, weathers dark brown, fossiliferous	36.0
5	Limestone, bluish-gray to purplish, fine-grained, weathers slabby, contains well-preserved and abundant *Lingula*	4.0
4	Shale, mostly covered	5.0
3	Limestone, bluish-gray, weathers brown, contains *Lingula* and brachiopods, slabby, fine-grained	3.0
2	Siltstone, yellowish-buff, unfossiliferous, shaly	48.0
1	Shale, buff, silty, partially covered	30.0
	Total Dinwoody	369.5
	Phosphoria formation	
	Limestone, black, silty, weathers buff, contains *Chonetes* and *Avonia*	75.0

13.—Snake River, Wyoming Range, Wyoming

½ mile east and 1½ miles south of junction of Martin Creek and Snake River NW.¼ secs. 16 and 21, T. 38 N., R. 116 W.

Bed No.		Feet
	Thaynes formation (section incomplete)	
88	Siltstone, gray, limy, weathers brown, some hundreds of feet to another massive limestone, all covered	35.0
87	Covered	25.0
86	Limestone, silty, weathers dark brown	38.0
85	Covered	55.0
84	Limestone, conglomeratic, contains miscellaneous pebbles	2.0
83	Covered	115.0
82	Limestone, gray, hard, dense, silty, weathers earthy	2.0
81	Siltstone, limy, red beds near top, partially covered	46.0
80	Siltstone, limy, weathers brown	13.0
79	Limestone, grayish-pink, silty	2.0
78	Siltstone, maroon, slabby, contains some limestone beds	14.0

Bed No.		Feet
77	Siltstone, shaly, red, partially covered	8.0
76	Siltstone, maroon, calcareous	1.0
75	Shale, maroon, silty	9.0
74	Siltstone, olive-gray, shaly	17.0
73	Limestone, gray, hard, dense, fine-grained	5.0
72	Siltstone, calcareous, mostly covered	21.0
71	Siltstone, blocky, calcareous	22.0
70	Limestone, cavernous	3.0
69	Limestone, punky and gray, silty	8.0
68	Siltstone, olive, silty, shaly	5.0
67	Limestone, punky, cavernous	2.5
66	Limestone, pink, silty	4.0
65	Shale, red, silty	5.0
64	Limestone, light-gray, hard, dense, weathers brown, lower part conglomeratic, with small limestone pebbles, upper part pink	13.0
63	Siltstone, shaly	7.0
62	Limestone, buff, silty, blue where fresh	6.0
61	Shale, olive-gray, silty	4.0
60	Limestone and platy olive siltstone, interbedded	16.0
59	Siltstone, olive, platy	4.0
58	Limestone, blue, silty, fine-grained, blocky, weathers buff	1.5
57	Siltstone, shaly, partially covered	9.0
56	Limestone, gray, silty	8.5
		526.5
	Woodside formation	
55	Shale, red, silty, contains a few thin red platy sandstone beds	200.0
54	Sandstone, red, fine-grained and interbedded red shales, makes ridge	135.0
53	Shale, red, soft	25.0
52	Sandstone, red, interbedded with red silty shales, fine-grained, 1 foot of gray sandstone at top	100.0
51	Shale, red, soft, partially covered	55.0
50	Siltstone, whitish-gray, calcareous	1.5
49	Shale, red, partially covered	115.0
48	Shale, gray, partially covered	50.0
	Total Woodside	681.5
	Dinwoody formation	
47	Siltstone, calcareous, hard, brittle, more limy and coarse-grained toward top, also platy	27.0
46	Siltstone, gray, calcareous, platy, weathers brown	20.0
45	Shale, olive, silty	5.0
44	Siltstone, buff, calcareous, platy and blocky, weathers brown	20.0
43	Siltstone, contains small pelecypods, partially covered	10.0
42	Siltstone, gray, weathers buff	25.0
41	Siltstone, mostly covered	10.0
40	Siltstone, buff and gray, hard, calcareous, brittle, weathers brown, contains small pelecypods	10.0
39	Shale, silty, partially covered	10.0
38	Siltstone, hard, calcareous, quartzitic, contains numerous *Claraia* and small pelecypods	25.0
37	Shale, olive, silty, platy	10.0
36	Limestone, gray, sandy, weathers brown, shaly toward top	6.0
35	Shale, olive, silty	1.0
34	Limestone, buff, sandy, contains no fossils	5.0
33	Shale, gray, silty	1.5
32	Limestone, buff, silty, sandy, contains few *Lingula*	2.0
31	Shale, olive, silty	8.0
30	Limestone, contains *Lingula* and silty shale, somewhat flaggy	20.0

Bed No.		Feet
29	Limestone, bluish-gray, fine-grained, some layers filled with *Lingula*, blocky, weathers brown	13.0
28	Shale, gray, argillaceous, silty	15.0
27	Limestone, bluish-gray, brittle, dense, contains no fossils	0.4
26	Shale, gray, argillaceous	5.0
25	Limestone, bluish-gray, fine-grained, blocky, abundant *Lingula* at top	4.0
24	Shale, gray, silty, several calcareous beds with *Lingula*	10.0
23	Limestone, shaly toward top, contains *Lingula*	6.0
22	Shale, gray, silty	5.0
21	Limestone, thin and unfossiliferous, shales, cephalopods, *Lingula*, *Myalina*	10.0
20	Shale, gray, silty	5.0
19	Limestone, contains *Lingula*	2.0
18	Shale, gray, silty	2.0
17	Limestone, silty, gray, weathers brown, contains *Lingula*	4.0
16	Shale, gray, silty, contains some thin siltstones	10.0
15	Limestone, contains *Lingula*	1.0
14	Shale, gray, silty, partially covered	2.0
13	Limestone, bluish-gray, contains pelecypods and *Lingula*	3.5
12	Shale, bluish-gray, argillaceous, contains thin calcareous streaks	14.0
11	Limestone, bluish-gray, even-bedded, fine-grained, contains *Lingula*	7.0
10	Shale, silty and shaly siltstone, unfossiliferous	15.0
9	Same as 7, with *Lingula*, somewhat shaly, top is coquina of *Lingula*	11.5
8	Shale, gray, silty	2.0
7	Limestone, dark-gray, silty, weathers brown, contains sparse *Lingula* at top	5.0
6	Shale, silty, partially covered	4.0
5	Limestone, dark reddish-brown, fossiliferous, weathers dark brown	0.5
4	Shale, silty, partially covered	3.0
3	Siltstone, brown, calcareous, hard, weathers into slabs, makes prominent ridge, no fossils found, numerous concretions weathering to limonite	15.0
2	Siltstone, buff-gray, shaly, weathers brown	30.0
1	Covered	10.0
	Total Dinwoody	430.4

Phosphoria formation

Limestone, bluish-gray, silty, hard, fine-grained, contains no fossils

14.—North Piney Creek, Wyoming Range, Wyoming

Sec. 7, T. 31 N., R. 114 W.

Bed No.		Feet
	Thaynes formation (section incomplete)	
94	Limestone, gray, lower half contains maroon bands and chert, contains no fossils	100.0
93	Siltstone, mostly covered	15.0
92	Limestone, maroon, sandy, contains red chert bands	4.0
91	Siltstone, blocky, calcareous, mostly covered, weathers buff to brown	30.0
90	Limestone, light-gray, fine-grained, contains no fossils	2.0
89	Siltstone, brown, calcareous, blocky	30.0
88	Limestone, gray to buff, conglomeratic, with siltstone pebbles	5.0
87	Limestone, maroon, massive, conglomeratic, with siltstone pebbles	10.0
		196.0
	Woodside formation	
86	Siltstone, red	65.0
85	Siltstone, red, poorly exposed	75.0
84	Shale and siltstone, red, silty	70.0
83	Shale, red, silty	13.0
82	Siltstone, red, shaly and platy	38.0

Bed No.		Feet
81	Siltstone red, shaly	37.0
80	Siltstone, gray, blocky	2.0
79	Siltstone, red, shaly, poorly exposed	45.0
78	Siltstone, red, makes two or three hogback ridges	50.0
77	Siltstone, gray to white	1.0
76	Siltstone, brown and red	22.0
75	Siltstone, red, shaly	27.0
74	Sandstone, fine-grained, interbedded with red shale	13.0
73	Shale, maroon	5.0
72	Sandstone, red, slabby, gray ledge at top	10.0
71	Shale, red	18.0
70	Sandstone, red, slabby	7.0
69	Shale, red, partly covered	63.0
68	Sandstone, red	3.0
67	Sandstone, red, coarse-grained, contains gray spots, has conglomeratic appearance	33.0
66	Shale, red, gypsiferous	13.0
65	Sandstone, red, shaly	23.0
64	Shale, red, silty, partly covered	33.0
63	Sandstone, brown and green	3.0
62	Shale, interbedded red and gray	1.0
61	Shale, red, silty	16.0
60	Shale, olive, silty	3.0
59	Shale, maroon	6.0
	Total Woodside	695.0
	Dinwoody formation	
58	Limestone, buff to maroon, shaly, weathers tan	14.0
57	Limestone, pink to blue, granular	2.0
56	Limestone, light-gray, massive, weathers light-brown	0.5
55	Shale, greenish, partly covered	4.5
54	Siltstone, olive	1.5
53	Shale, mostly covered	6.0
52	Limestone, gray, blocky, silty	5.0
51	Same as 50 but less massive	24.0
50	Siltstone, olive to buff, massive, blocky, rippled	25.0
49	Shale, green, platy	5.0
48	Limestone, buff, silty	3.0
47	Siltstone, calcareous, blocky	7.0
46	Limestone, blue, contains some chert bands	4.0
45	Siltstone, olive, calcareous, massive, storm rollers at base, contains *Claraia*	3.0
44	Limestone, bluish-gray, irregular bedding	1.0
43	Siltstone, olive, platy	4.0
42	Limestone, blue, and interbedded olive siltstone	1.5
41	Siltstone, olive, platy	2.5
40	Limestone, blue, contains some interbedded siltstone	4.0
39	Covered	25.0
38	Siltstone, calcareous, blocky at top, partly covered	35.0
37	Same as 36 but less massive and blocky	5.0
36	Siltstone, calcareous, weathers tan	11.0
35	Limestone, blue	0.3
34	Limestone, silty, blocky, weathers brown	4.5
33	Limestone, blue	0.3
32	Limestone, silty, blocky, weathers brown	4.0
31	Limestone, blue	1.0
30	Limestone, silty, blocky, weathers brown	2.0
29	Limestone, blue	0.5

Bed No.		Feet
28	Siltstone, green, shaly	19.0
27	Limestone, blue, and interbedded siltstone	2.0
26	Siltstone, blocky, calcareous, weathers brown, contains a few thin obscure limestones	16.0
25	Limestone, blue, hard, fine-grained, dense	1.0
24	Siltstone, rippled, contains a few thin limestones	11.0
23	Limestone, blue, contains *Lingula*	3.0
22	Siltstone, shaly	4.0
21	Limestone, blue, contains *Lingula*	2.0
20	Limestone, contains a few interbedded siltstones and *Lingula*	3.0
19	Siltstone, olive-gray, shaly	28.0
18	Limestone, blue, contains *Lingula*	1.5
17	Siltstone, olive-gray, shaly	18.0
16	Limestone, blue	1.0
15	Siltstone, shaly	2.5
14	Limestone, blue, contains *Lingula*	1.5
13	Siltstone, poorly exposed	5.0
12	Limestone, interbedded with green siltstone	5.0
11	Siltstone, shaly	4.5
10	Limestone, interbedded with siltstone	2.5
9	Siltstone, blue-gray, shaly, dense	8.0
8	Limestone, blue, contains some interbedded siltstone	1.5
7	Siltstone, olive-gray, interbedded with olive limestones, contains *Lingula*	7.0
6	Limestone, blue-gray, contains *Lingula*, hard, fine-grained, dense	4.0
5	Siltstone, blue-gray, calcareous, thin-bedded	11.0
4	Shale, mostly covered	8.0
3	Siltstone, olive-gray, shaly	11.0
2	Siltstone, olive-gray, slabby	15.0
1	Siltstone, greenish-gray, platy, calcareous, mostly covered	40.0
	Total Dinwoody	442.1
	Phosphoria formation	
	Shale, black	1.0
	Limestone, gray, silty, contains black chert	

15.—MacDougals Pass, Salt River Range, Wyoming

S. center sec. 9, T. 33 N., R. 117 W.

Bed No.		Feet
	Thaynes formation (section incomplete)	
100	Limestone, blue and brown, thin-bedded, *Meekoceras* zone at top	25.0
99	Limestone, gray, granular	20.0
98	Limestone, pink, granular	3.0
97	Limestone, gray, granular	30.0
96	Limestone, pink, granular	20.0
95	Limestone, blue, granular, many silicified pelecypods present	15.0
94	Limestone, bluish-gray, granular, fossiliferous, contains pelecypods	4.0
93	Limestone, gray, silty or calcareous siltstone	23.0
		140.0
	Woodside formation	
92	Siltstone, gray to buff, calcareous	21.0
91	Siltstone, red, platy	15.0
90	Siltstone, gray, platy	12.0
89	Sandstone, red, fine, shaly	225.0
88	Shale, red, silty, poorly exposed	100.0
87	Siltstone, olive, hard, platy	14.0
86	Limestone, drab-buff, granular, no fossils	1.0

Bed No.		Feet
85	Siltstone, olive, hard, platy	11.0
84	Shale, red, silty, poorly exposed	40.0
83	Siltstone, red, blocky, hard	15.0
82	Siltstone, red and gray, interbedded, shaly	15.0
81	Shale, red, silty	25.0
80	Shale, greenish-gray, silty	1.0
79	Siltstone, red, platy, and red shale	14.0
78	Shale, turquoise-colored, olive, platy	1.0
77	Shale, red, platy, silty	19.0
76	Siltstone, light-gray and red, interbedded	1.5
75	Shale, red, silty	98.0
74	Limestone, whitish-gray, sandy	0.5
73	Shale, red, silty	3.0
72	Siltstone, red and gray, platy	3.0
71	Shale, brick-red	6.0
70	Shale, light-gray, silty	1.0
69	Shale, brick-red, silty	15.0
68	Siltstone, olive, platy	4.0
67	Shale, dark brick-red, silty	7.0
	Total Woodside	668.0
	Dinwoody formation	
66	Siltstone, olive	11.0
65	Siltstone, calcareous, and olive silty shale	32.0
64	Covered	20.0
63	Sandstone, light-gray, calcareous, conglomeratic pebbles of limestone	20.0
62	Limestone, blue, dense, massive, relatively unfossiliferous, consists of fine layers of dark-blue and gray limestone, cross-bedded on small scale, contains silicified fossils	5.0
61-b	Siltstone, calcareous, massive, weathers brown	5.0
61-a	Limestone, blue, dense, contains silicified fossils	1.0
60	Siltstone, calcareous and interbedded shaly limestone	13.0
59	Limestone, blue, dense, contains myriads of silicified pelecypods	1.5
58	Siltstone, buff, calcareous	37.0
57	Siltstone, massive, contains lenses of chert and dense blue limestone	15.0
56	Shale, olive, platy	2.0
55	Limestone, blue, dense, hard, fossiliferous	1.0
54	Siltstone, calcareous, massive, heavy-bedded	28.0
53	Siltstone, olive and maroon, thin-bedded, shaly	7.0
52	Siltstone, olive-gray, limy, rippled, massive in upper parts	10.0
51	Limestone, blue, hard, dense, contains many small fossils including gastropods, appears to be oölitic in parts	1.5
50	Shale and siltstone, olive, chocolate-maroon, silty	25.0
49	Shale, olive, silty	10.0
48	Limestone, gray, hard, weathers gray, fossiliferous	3.5
47	Shale and siltstone, some rippled sandstone, fossiliferous, appears to have mud cracks	13.0
46	Siltstone, olive to buff, shaly	1.5
45	Shale, gray, silty	3.5
44	Siltstone, olive, shaly, weathers brown, interbedded with bluish-gray limestone	3.0
43	Shale, silty, and gray shaly siltstones	14.0
42	Limestone, gray, hard, weathers gray, fine-grained, contains echinoid spines	0.5
41	Siltstone, olive to buff, flaggy, calcareous	1.0
40	Limestone, bluish-gray	1.0
39	Shales, silty, and olive, chocolate, platy siltstone	20.0

Bed No.		Feet
38	Limestone, bluish-gray, and interbedded siltstone	2.5
37	Shale, buff to brown, with very fossiliferous leached siltstone, contains pelecypods, *Lingula*, and *Mentzelia?*	30.0
36	Limestone, crystalline, fossiliferous and interbedded siltstone, contains echinoid spines, *Lingula*, pelecypods	2.0
35	Shale, silty, and gray interbedded siltstone, weathers brown, contains *Claraia*	10.0
34	Shale, olive, chocolate	10.0
33	Siltstone, greenish-gray, weathers brown	1.0
32	Limestone, gray, hard, weathers gray, contains *Mentzelia?*, ammonoids, *Claraia*	7.5
31	Siltstone, greenish-buff, hard, resistant, shale at top	5.0
30	Siltstone, shaly	19.0
29	Limestone, gray, hard, weathers gray, fossiliferous	0.5
28	Siltstone, gray, limy	3.5
27	Shale, olive-gray, silty	11.0
26	Shale, gray, silty, 1½-foot limestone at top, partly covered	5.0
25	Limestone, gray, weathers light gray, contains *Lingula*	4.0
24	Shale, covered	6.0
23	Limestone, gray, weathers gray, contains *Lingula*	1.0
22	Shale, gray, silty, partly covered	5.0
21	Limestone and interbedded siltstone, weathers brown, contains *Lingula*	4.0
20	Siltstone, olive-gray, platy, limy, weathers brown, contains *Lingula*	9.0
19	Limestone, gray, hard, and interbedded siltstone, contains large pelecypods and *Lingula*	16.0
18	Shale, olive-gray, silty, thin siltstone bands, contains *Lingula*	20.0
17	Shale, olive-gray, silty	4.5
16	Limestone, contains *Lingula*	0.2
15	Shale, gray, silty, partly covered	2.5
14	Limestone, bluish-gray, fine-grained, weathers brown, slabby, contains *Lingula*	3.5
13	Shale, buff, soft, mostly covered	16.0
12	Limestone, bluish-gray, platy, and interbedded buff siltstone, contains *Lingula* and large pelecypods	4.0
11	Shale, olive and chocolate, silty	14.0
10	Siltstone, olive, shaly, contains *Lingula*	2.0
9	Shale, olive-gray, soft	14.0
8	Limestone, gray-blue, fine-grained, weathers brown, abundant *Lingula*	2.0
7	Shale, gray, silty	5.0
6	Limestone, gray, silty, weathers brown	0.5
5	Shale, gray, silty	8.0
4	Limestone, with interbedded siltstone, contains *Lingula*	9.0
3	Shale, gray, silty, partly covered	5.0
2	Siltstone, gray, platy, calcareous, weathers brown to buff, contains *Lingula* in top bed	87.0
1	Siltstone, gray, limy, slabby, weathers brown to gray, in normal outcrop forms slope, few hard ledges projecting	80.0
	Total Dinwoody	735.2

Phosphoria formation

Limestone, black, phosphatic, platy, contains pyrite concretions	11.0
Shale, black, phosphatic and limy	12.5
Limestone, phosphatic, nodular, black	1.0
Limestone, phosphatic, black, massive, nodular in bands	3.0
Phosphate, gray, low grade, speckled	4.0
Covered	4.0

Rex chert, top.

16.—Swift Creek Canyon, Salt River Range, Wyoming, Just East of Afton
S. E. ¼ sec. 19, T. 32 N., R. 118 W.

Bed No.		Feet
	Thaynes formation (section incomplete)	
110	Shale, dark gray to black, silty, weathers olive-buff	50.0
109	Limestone, gray, fine-grained, weathers brown, *Meekoceras* zone begins 12 feet above base	26.0
108	Siltstone, brown	15.0
107	Limestone, blue, fine-grained, weathers brown	1.0
106	Siltstone, platy, weathers brown	7.0
105	Limestone, bluish, fine-grained, weathers brown	1.0
104	Siltstone, shaly	6.0
103	Limestone, bluish, fine-grained, weathers brown	5.0
102	Siltstone, brown, blocky	11.0
101	Limestone, light gray, massive, fine-grained, makes very prominent ridge, contains small brachiopods (terebratuloids)	25.0
100	Shale, silty	10.0
99	Limestone, light gray, forms prominent ridge	20.0
		177.0
	Woodside formation	
98	Covered	15.0
97	Siltstone, brown, blocky, quartzitic	15.0
96	Covered	15.0
95	Shale, red, silty, partially covered, contains few thin red sandstone bands	320.0
94	Sandstone, red, thin-bedded	3.0
93	Shale, red, silty	30.0
92	Siltstone, greenish-gray, shaly, slabby	50.0
91	Siltstone, maroon, slabby, fine-grained	1.5
90	Siltstone, greenish-gray, slabby, brittle	6.0
89	Shale, red, silty, with few thin red and gray sandstone bands	210.0
88	Limestone, silty, whitish-gray	2.0
87	Shale, red, silty, poorly exposed	14.0
86	Limestone, maroon, fine-grained	0.5
85	Shale, red, silty, contains few thin red sandstone bands	10.0
84	Shale, gray, silty	4.0
83	Shale, red, silty	6.5
82	Shale, olive, green, silty, platy	5.0
	Total Woodside	707.5
	Dinwoody formation	
81	Limestone, light gray, brittle, weathers tan	2.0
80	Siltstone, olive, platy	2.0
79	Limestone, light gray, weathers gray, brittle	17.0
78	Siltstone, blue, calcareous, contains chert nodules, blocky, brittle, weathers brown	19.0
77	Sandstone, light-gray, almost quartzitic, forms low scarp	33.0
76	Limestone, blue, silty, striped, fossiliferous	11.0
75	Siltstone, calcareous, blocky, weathers brown	15.0
74	Limestone, blue, siliceous, fine-grained, fossiliferous	9.0
73	Siltstone, blocky, calcareous	11.0
72	Limestone, blue, silty, contains color bands and pebbles	4.0
71	Siltstone, very calcareous, quartzitic, massive, blue where fresh, weathers brown	20.0
70	Limestone, blue	1.0
69	Siltstone, olive, platy	5.0
68	Limestone, blue	0.2
67	Siltstone, olive, platy	5.0

Bed No.		Feet
66	Siltstone, blue where fresh, massive, quartzitic	7.0
65	Shale, silty	4.0
64	Limestone	1.0
63	Siltstone, platy, contains few thin limestone	30.0
62	Limestone and siltstone, interbedded	3.0
61	Siltstone, olive, shaly	17.0
60	Siltstone, olive, hard, platy	4.0
59	Siltstone, olive and gray limestone interbedded	6.0
58	Siltstone, olive, platy	15.0
57	Limestone, bluish-gray, weathers pink	2.0
56	Siltstone, olive, calcareous, platy	15.0
55	Limestone, gray, fine-grained	3.5
54	Shale, chocolate and olive, silty	7.5
53	Shale, olive, platy	1.5
52	Limestone, gray, granular	2.0
51	Shale, chocolate, olive, silty	13.0
50	Limestone, gray, granular	1.0
49	Siltstone, calcareous	2.0
48	Limestone, blue, granular, and siltstone	1.0
47	Siltstone, hard, calcareous, platy, contains *Lingula*	15.0
46	Limestone, blue, granular, even-bedded	2.0
45	Siltstone, olive, platy	16.0
44	Limestone, blue	2.0
43	Siltstone, very calcareous, contains one or two 2-inch limestone bands	8.0
42	Limestone, blue, fossiliferous	2.0
41	Siltstone, calcareous, contains *Lingula*	2.5
40	Limestone	0.5
39	Siltstone, olive, shaly	1.5
38	Limestone, bluish, dense	8.0
37	Shale, platy, silty	10.0
36	Limestone, blue, dense, granular	9.0
35	Shale, olive, silty	6.0
34	Limestone, blue, dense	1.5
33	Shale, olive, silty	6.0
32	Limestone, cephalopods seen, contains *Lingula*	4.0
31	Shale, chocolate and olive, silty	2.0
30	Limestone, blue, dense, contains *Lingula*	0.3
29	Shale, chocolate and olive, silty	3.5
28	Limestone, gray, some silty bands, contains *Lingula*	20.0
27	Siltstone, shaly	19.0
26	Limestone, silty	0.2
25	Siltstone, shaly	1.0
24	Limestone, blue, dense, contains *Lingula*	2.5
23	Shale, silty, contains two or three thin *Lingula* limestones	9.5
22	Limestone, blue, dense, contains *Lingula*	0.3
21	Shale, silty, contains three or four thin *Lingula* limestone beds	4.5
20	Limestone, blue, contains *Lingula*, interbedded with siltstone	4.0
19	Shale, olive, silty, contains two or three 1-inch *Lingula* limestone beds	10.0
18	Limestone, blue, dense, contains *Lingula*	0.3
17	Shale, olive, silty	5.0
16	Siltstone, gray, slabby	2.5
15	Shale, gray, silty	5.0
14	Limestone, blue, dense, contains *Lingula*	0.5
13	Shale, gray, silty	4.0
12	Limestone, blue, dense, contains *Lingula*	1.0
11	Shale, gray	1.5
10	Limestone, blue, silty, weathers buff, contains *Lingula*	1.5
9	Shale, gray, silty	5.0

Bed No.		Feet
8	Limestone, blue, dense, contains *Lingula*	3.0
7	Siltstone, shaly, slabby	7.0
6	Limestone, blue, dense, contains *Lingula*	6.0
5	Siltstone, gray, shaly	8.5
4	Siltstone, weathers brown, upper part slabby, calcareous	14.0
3	Siltstone, buff, platy	15.0
2	Siltstone, platy	83.0
1	Covered	37.0
	Total Dinwoody	660.3

Phosphoria formation

Rex chert

17.—Sublette Mountains, Wyoming—Canyon 3 Miles North of Raymond Canyon

Near center of E. ½ and center of NE. ¼ sec. 19, T. 27 N., R. 119 W.

Bed No.		Feet
	Thaynes formation (section incomplete)	
75	Limestone, gray, heavy-bedded, granular, *Meekoceras* zone, becomes brown above	55.0
74	Limestone and siltstone, interbedded	19.0
		74.0
	Woodside formation	
73	Covered, probably contains no red shale	40.0
72	Shale, red, some olive streaks, partially covered	35.0
71	Siltstone, olive, platy	5.0
70	Siltstone, chocolate, platy	25.0
69	Siltstone, olive, platy	9.0
68	Shale, chocolate and maroon, silty	10.0
67	Siltstone, olive and chocolate, platy, rippled	6.0
66	Sandstone, chocolate and maroon, shale	67.0
65	Shale, dark brick-red to maroon, mostly covered	30.0
64	Siltstone, calcareous, blocky, greenish, some silty limestone with poorly preserved pelecypods	22.0
63	Siltstone, olive and chocolate, slabby	20.0
62	Shale, red to maroon, silty, poorly exposed	180.0
61	Siltstone, olive, platy, poorly exposed	18.0
60	Shale, largely red to maroon, partially covered	50.0
59	Shale, buff, and siltstone, partially covered	50.0
	Total Woodside	567.0
	Dinwoody formation	
58	Limestone, light-gray, sandy	30.0
57	Sandstone, light gray to buff, calcareous, grades upward into sandy limestone	30.0
56	Limestone, blue, fine-grained, mottled and striped, cross-bedded on small scale	24.0
55	Limestone (?), covered	10.0
54	Siltstone, calcareous, platy or silty limestone	10.0
53	Limestone, blue, fine-grained, mottled and striped	5.0
52	Limestone, blue, silty, interbedded with siltstone	5.0
51	Siltstone, becoming massive and brown in upper part, mostly covered	29.0
50	Limestone, bluish-gray, coarse-grained, fossiliferous	2.0
49	Siltstone, buff to brown, massive, calcareous	19.0
48	Limestone, interbedded with platy siltstone	4.0
47	Siltstone, platy, calcareous, contains poorly preserved *Lingula*	26.0

Bed No.		Feet
46	Limestone, bluish-gray, fine-grained	2.0
45	Siltstone, platy	15.0
44	Limestone, bluish-gray	1.0
43	Siltstone, interbedded with bluish limestone	3.5
42	Limestone, bluish-gray, fine-grained	2.0
41	Siltstone, buff, platy	12.0
40	Limestone, bluish-gray, fine-grained	1.0
39	Siltstone, buff to brown, massive, calcareous	3.0
38	Shale, buff, silty	10.0
37	Limestone, bluish-gray, fine-grained, weathers brown	1.5
36	Covered	14.0
35	Limestone, bluish-gray, and interbedded shales	2.0
34	Siltstone, buff to brown, calcareous, blocky	1.0
33	Siltstone, calcareous, blocky, interbedded with silty shales	8.0
32	Siltstone, olive to buff, slabby	5.0
31	Siltstone, buff, platy, and silty shale	12.0
30	Siltstone, gray to buff, slabby	3.0
29	Shale, buff, silty	5.5
28	Limestone, bluish-gray, granular, contains brachiopod similar to *Mentzelia?*	3.0
27	Shale, olive, silty	5.0
26	Limestone, blue, dense, contains *Lingula,* interbedded with gray silty shales	8.0
25	Shale, gray, silty	3.0
24	Limestone, interbedded with siltstone	1.0
23	Shale, gray, silty	10.0
22	Limestone, bluish, fine-grained, weathers brown, contains *Lingula*	3.0
21	Shale, buff, silty	4.5
20	Limestone, blue, dense, contains *Lingula,* interbedded with buff silty shale	4.0
19	Shale, buff, silty	4.0
18	Limestone, bluish, dense, contains *Lingula*	0.8
17	Shale, buff, silty	8.5
16	Limestone, bluish, weathers brown, contains *Lingula,* interbedded with siltstone	3.0
15	Shale, buff, silty, mostly covered	10.0
14	Siltstone, buff, slabby, weathers brown, contains *Lingula*	2.0
13	Shale, gray, silty	15.0
12	Limestone, blue, fine-grained, weathers brown, contains *Lingula,* interbedded with buff siltstone	5.0
11	Shale, gray, silty	15.0
10	Siltstone, buff, calcareous, contains abundant *Lingula,* upper foot is blue limestone	5.5
9	Shale, buff to olive, platy, silty	10.0
8	Limestone, bluish-gray, fine-grained, platy, contains *Lingula*	1.0
7	Shale, olive-buff, silty	1.5
6	Siltstone, brown, shaly	6.0
5	Shale, buff, silty	16.0
4	Limestone, bluish-gray, contains *Lingula,* interbedded with thin shales	4.5
3	Shale, greenish-gray, silty, contains a few poorly preserved clams	7.0
2	Siltstone, buff, platy and slabby, partly covered, upper 15 feet becomes blocky and calcareous, weathers brown	115.0
1	Covered	8.0
	Total Dinwoody	574.8

Phosphoria formation

Shale, black, carbonaceous, contains a few phosphate nodules 13.0
Rex chert, forms scarp.

18.—Cokeville, Wyoming, Wyoming Range

¾ mile east from U. S. Highway 30 by the Stauffer Chemical Company's phosphate mine, T. 24N., R. 119W.

Bed No.		Feet
	Thaynes formation (section incomplete)	
92	Shale, brown, silty, calcareous bands, *Meekoceras* zone at base	100.0±
91	Limestone, pink and gray, granular	70.0
90	Limestone, bluish-gray, granular	11.0
89	Limestone, buff to brown, silty, poorly exposed	17.0
		198.0
	Woodside formation	
88	Siltstone, gray, platy	10.0
87	Shale, maroon, silty	15.0
86	Shale, olive, silty, contains some siltstone	11.0
85	Shale, maroon, silty and sandy	80.0
84	Sandstone, maroon	2.0
83	Shale, maroon, silty	25.0
82	Shale, gray, silty	12.0
81	Shale, maroon, and siltstones, partially covered	284.0
80	Sandstone, maroon, fine-grained, slabby	5.0
79	Shale, gray, silty	3.5
78	Limestone, light-gray	3.0
77	Siltstone, gray, calcareous, blocky	3.5
76	Sandstone, gray, fine-grained	3.5
75	Shale, maroon, silty	10.0
74	Shale, olive, silty	2.5
73	Siltstone, buff, even, 1-inch cavernous limestone at top, fossiliferous	5.0
72	Shale, maroon, silty, poorly exposed	14.0
	Total Woodside	489.0
	Dinwoody formation	
71	Limestone, blue-gray, fine-grained, weathers buff	9.0
70	Siltstone, yellowish and olive, rippled, thin-bedded	9.0
69	Limestone, dark-gray, silty, even-bedded, weathers buff, contains no fossils	20.0
68	Limestone, silty, blocky, unfossiliferous, purplish near top	5.0
67	Siltstone, buff, platy	5.0
66	Limestone, bluish-gray, silty, weathers buff, weathers into small angular pieces, contains few fossils	40.0
65	Shale, olive, silty	3.5
64	Limestone, gray to pink, silty	3.0
63	Limestone and interbedded olive shale	3.0
62	Shale, olive, silty	3.5
61	Limestone, pink, granular and interbedded olive and silty shale	4.5
60	Limestone or limy siltstone, brown, blocky, contains *Lingula* and pelecypods	7.0
59	Siltstone, olive, shaly	10.0
58	Limestone, pink to maroon, granular, contains abundant pelecypods	3.0
57	Shale, chocolate to maroon, with olive streaks	4.0
56	Shale, olive, silty with interbedded platy siltstone	8.0
55	Limestone, gray, silty, weathers brown to black, contains *Myalina*	2.0
54	Siltstone, shaly, contains a few thin limestones	10.0
53	Siltstone, olive, shaly	20.0
52	Limestone, gray	0.2
51	Siltstone, buff, shaly, interbedded with olive shale, contains large *Lingula*	4.0
50	Limestone, bluish-gray, fine-grained	0.3
49	Siltstone, olive, shaly, contains a few 1-inch limestone beds	20.0
48	Limestone, gray, granular	1.0

Bed No.		Feet
47	Shale, olive, silty and siltstone	3.5
46	Siltstone, greenish, calcareous, interbedded with thin blue limestone	9.5
45	Shale, gray and chocolate, silty, contains *Lingula*	8.0
44	Limestone, bluish-gray, fine-grained, some pinkish layers	4.0
43	Shale, silty, poorly exposed, contains *Lingula*	9.0
42	Limestone, bluish-gray, granular, interbedded with shale	5.0
41	Shale, olive, silty, contains some chocolate streaks	3.0
40	Limestone, bluish-gray, granular	1.0
39	Shale, green, silty	2.0
38	Siltstone, olive	2.0
37	Shale, olive, silty	2.0
36	Limestone, bluish-gray, fine-grained	1.5
35	Siltstone, olive, contains some chocolate streaks	11.0
34	Limestone, bluish-gray, thin-bedded, weathers yellowish	2.0
33	Shale, mostly covered	11.0
32	Limestone, bluish, granular	1.0
31	Shale, gray, silty	3.0
30	Limestone, bluish-gray, fine-grained	1.0
29	Shale, partially covered	12.0
28	Limestone, bluish-gray, fine-grained, weathers brown, contains *Lingula*	2.0
27	Shale, partially covered	7.0
26	Limestone, bluish-gray, fine-grained	1.5
25	Shale, buff, silty	3.0
24	Limestone, bluish-gray, fine-grained	3.0
23	Shale, buff, silty	6.0
22	Limestone, bluish-gray, contains *Lingula*	3.0
21	Shale, mostly covered	25.0
20	Limestone, gray, hard, weathers brown	0.5
19	Shale, gray, silty	13.0
18	Limestone, poorly exposed	2.5
17	Shale, gray, silty	3.0
16	Limestone, blue, contains *Lingula*	1.0
15	Shale, partially covered	14.0
14	Limestone, bluish-gray, thin-bedded, contains *Lingula*	4.0
13	Shale, buff, silty	12.5
12	Limestone, bluish-gray with interbedded shales, contains *Lingula*	3.5
11	Siltstone, shaly	6.0
10	Limestone, silty, thin-bedded	0.5
9	Shale, gray, silty, platy	16.0
8	Limestone, bluish-gray, hard, contains shale partings, ammonoids and Bellerophons	4.5
7	Shale, gray, silty	1.0
6	Limestone, blue with interbedded siltstone, contains *Lingula*	1.5
5	Siltstone, buff, shaly	5.0
4	Siltstone, brown, limy, blocky	21.0
3	Siltstone, buff and olive, platy	80.0
2	Siltstone, buff, platy	9.0
1	Siltstone, buff, platy	7.0
	Total Dinwoody	542.5

Phosphoria formation

Shale, black, carbonaceous	7.0
Shale, limy, carbonaceous	1.0
Limestone, black, carbonaceous	2.0
Shale, black, carbonaceous	2.5
Covered, contains some silty beds	125.0
Rex chert, forms prominent ridge	....

19.—Montpelier Canyon, 1½ Miles East of Montpelier, Idaho
Sec. 31, T.27N., R.45E.

Bed No.		Feet
	Thaynes formation (section incomplete)	
50	Shale, mostly covered	....
49	Limestone, gray and buff, thin-bedded, *Meekoceras* zone above middle	70.0
		70.0
	Woodside formation	
48	Shale, mostly covered	80.0
47	Siltstone, brown, flaggy, interbedded with even-bedded gray limestone	30.0
46	Limestone, gray, sugary	9.0
45	Siltstone, maroon and green, interbedded with gray unfossiliferous limestone	50.0
44	Siltstone, green and purple, platy	11.0
43	Limestone, gray, subgranular, contains pelecypods	4.0
42	Siltstone, greenish, platy	72.0
41	Limestone, dark-gray to purple, fine-grained, sparsely fossiliferous	4.0
40	Shale, partly maroon, partially covered	110.0
39	Limestone, purplish, fine-grained, unfossiliferous, conglomeratic, with olive siltstone pebbles	0.4
38	Shale, maroon	22.0
	Total Woodside	392.4
	Dinwoody formation	
37	Siltstone, hard, platy, and a little interbedded gray silty limestone, unfossiliferous	19.0
36	Limestone, gray, fine-grained, unfossiliferous, silty	1.5
35	Shale and siltstone, olive, some maroon(?)	42.0
34	Limestone, and gray quartzitic siltstone, relatively unfossiliferous	19.0
33	Siltstone, olive, hard	16.0
32	Limestone, gray, blocky, contains few fossils	3.5
31	Shale, mostly covered	10.5
30	Limestone, gray, fine-grained, massive, conglomeratic at base with siltstone pebbles, contains few fossils	11.0
29	Limestone, gray, fine-grained, and olive hard siltstone, partly covered; there may be a little shale present	34.0
28	Sandstone, olive to buff, cross-bedded and rippled, hard, fine-grained, contains a few thin beds of limestone	57.0
27	Limestone, gray, fine-grained	7.5
26	Sandstone or siltstone, brown	19.5
25	Limestone, bluish-gray, fine-grained, interbedded and gradational into fine sandstone	11.0
24	Siltstone, olive-buff, platy, makes ridge, rippled, contains some edgewise conglomerate	46.0
23	Limestone, blue, fine-grained	2.0
22	Shale, olive, mostly covered, silty, olive siltstone toward top	80.0
21	Limestone, gray, silty and interbedded siltstone, unfossiliferous, current ripple marks	3.0
20	Shale, olive-drab, silty	16.0
19	Limestone, drab below, blue above, silty, platy, contains *Myalina*	4.0
18	Shale, olive-drab	16.0
17	Limestone, drab, silty, unfossiliferous	4.0
16	Shale, olive, platy	5.0
15	Limestone, bluish-gray, sugary, coarsely veined with white calcite, contains few fossils	1.5
14	Siltstone, olive-drab, platy, calcareous	5.0
13	Shale, mostly covered	20.0
12	Limestone, bluish, fine-grained, even-bedded, contains brachiopods and a cephalopod found in float	2.0

Bed No.		Feet
11	Siltstone and shale, olive-gray, platy, hard	50.0
10	Shale, olive, platy, contains a few *Lingula*	30.0
9	Siltstone, hard, drab, contains *Lingula*	3.0
8	Shale, buff to olive-drab, silty, platy	42.0
7	Limestone, blue, fine-grained, contains small poorly preserved ammonites, weathers brown	1.0
6	Shale, gray, platy	2.0
5	Limestone, thin-bedded, like 3	0.3
4	Shale, gray, platy	4.0
3	Limestone, blue, weathers brown, fine-grained, unfossiliferous	0.3
2	Siltstone and olive shale, buff, platy, hard, contains pectinoids at base and possibly higher	23.0
1	Shale, gray, silty, weathers buff, platy, contains a few thin hard siltstone beds toward top, imprints of small cephalopods (ammonites) 5 feet from base, also contains rare *Lingula*	107.0
	Total Dinwoody	718.6

Phosphoria formation

20.—Section on North Side of Trapper Creek—9 Miles West of Melrose, Montana
W. part of sec. 22, T.2S., R.10W.

Bed No.		Feet
	Kootenai formation (section incomplete)	
35	Sandstone, heavy brown, with large quartz pebbles	30.0±
34	Sandstone, buff, platy, fine-grained	3.0±
33	Shale, greenish, silty, platy	2.0±
		35.0
	Dinwoody formation	
32	Limestone, light gray to light buff, silty, platy	6.0
31	Limestone, gray, banded, even-bedded, a few thin beds of platy silty shale	32.0
30	Covered (sandy shale)	20.0
29	Limestone, blue, silty, with included calcite crystals, weathers brown	2.0
28	Shale, platy, sandy, and platy buff sandstone	15.0
27	Limestone, blocky, fine-grained, weathers brown in places, elsewhere steel gray	9.0
26	Shale, buff, sandy, poorly exposed	5.0
25	Limestone, gray, hard	2.0
24	Covered, probably platy sandy shale	20.0
23	Limestone, steel gray, brittle, in thin beds	3.0
22	Covered, probably platy shaly sandstone	40.0
21	Sandstone, buff to olive, calcareous, hard, fine-grained	10.0
20	Limestone, steel gray, fine-grained, brittle	2.0
19	Shale, sandy, platy, poorly exposed	24.0
18	Limestone, steel gray, fine-grained, even beds about ¾ feet thick	5.0
17	Covered (shale?)	8.0
16	Limestone, steel gray, fine-grained, even-bedded	2.5
15	Shale, buff, fine-grained, sandy, platy, unfossiliferous	17.0
14	Limestone, steel gray to bluish, very fossiliferous, brittle, fine-grained, in even beds 1 inch thick, lower 2 feet brown with edgewise conglomerate, some platy calcareous sandstone	55.0
13	Sandstone, brown, calcareous, *Lingula* and *Myalina* present	30.0
12	Sandstone, buff, platy, and shale containing *Lingula*	30.0
11	Shale, gray, contains a thin *Lingula*-bearing limestone	30.0
10	Shale, gray, with a few *Lingula* beds, thin-bedded	30.0
9	Limestone, gray, weathers brown, contains *Lingula*	6.0
8	Shale, gray, with a few brown *Lingula* beds	6.0
7	Limestone, brown, contains *Lingula*	6.0

Bed No.		Feet
6	Shale, gray, silty, and interbedded brown *Lingula* siltstone, thin-bedded	17.0
5	Limestone, gray, weathers brown, platy, contains *Lingula*	5.0
4	Shale, gray, silty, interbedded with brown thin siltstone, contains *Lingula* and cephalopods	40.0
3	Siltstone, brown, platy, shaly, 1 foot blue limestone containing *Lingula* at top, weathers brown	11.0
2	Shale, bluish to gray, silty, unfossiliferous	100.0
1	Covered	50.0±
	Total Dinwoody	638.5
	Phosphoria formation	
	Rex, black, quartzite	

21.—Bear Creek, Reservoir Mountain, near Henry, Idaho
Four miles west of Blackfoot River Reservoir
Sec. 7, T. 6 S., R. 41 E.

Bed No.		Feet
	Thaynes formation (section incomplete)	
119	Limestone, light-gray, granular, cephalopod zone, *Meekoceras*	0.0±
	Woodside-Dinwoody formation, undifferentiated	
118	Shale, covered	85.0
117	Limestone, blue, fine-grained, contains brachiopods	24.0
116	Shale, buff	12.0
115	Limestone, blue, fine-grained	1.0
114	Shale, buff	17.0
113	Shale, covered	25.0
112	Limestone, blue, fine-grained, even-bedded	6.0
111	Sandstone and shale, brown, blocky	10.0
110	Shale, buff	50.0
109	Limestone, blue, fine-grained, contains brachiopods and some silty beds	45.0±
108	Shale, buff	100.0
107	Limestone, blue, poorly exposed	12.0±
106	Covered (shale?)	10.0
105	Limestone, blue, interbedded with brown siltstone	48.0
104	Shale This section is separated from the lower beds by a fault of undetermined throw. Beds 104 to 119 inclusive are overturned.	35.0±
103	Limestone, bluish-gray, filled with brachiopods and pelecypods	18.0
102	Shale, covered, contains *Terebratula margaritowi* and blocky brown sandstone	19.5
101	Limestone, gray, fine-grained, filled with brachiopods	7.0
100	Covered (shale?)	15.0
99	Sandstone, brown, blocky, lower half covered	65.0
98	Limestone, bluish-gray, filled with brachiopods, poorly exposed	5.0
97	Shale, covered	9.0
96	Limestone, bluish, fine-grained, contains crinoid stems (*Pentacrinus?*)	5.0
95	Shale, covered	15.0
94	Limestone, bluish-gray, shaly, filled with brachiopods	10.0
93	Shale and sandstone, covered	15.0
92	Limestone, bluish-gray, platy, fine-grained, filled with brachiopods, contact not well shown, *Myalina spathi, Terebratula margaritowi*	33.0
91	Shale, buff, silty, with a thin, blocky, fine sandstone near middle	30.0
90	Limestone, gray, fine-grained, platy, contains *Terebratula margaritowi*	32.0

Bed No.		Feet
89	Shale, covered	9.0
88	Limestone, gray, thin-bedded, contains some thin siltstone beds	54.0
87	Shale, covered	24.0
86	Siltstone, brown below, grading upward into thin-bedded gray limestone	49.0
85	Limestone, gray, even-bedded, fine-grained	4.0
84	Shale, buff, silty	27.5
83	Limestone, gray below, brown above, fine-grained, blocky	3.0
82	Shale, buff, silty	23.0
81	Limestone, bluish-gray, platy, fine-grained	2.0
80	Shale, covered, probably contains some limestone	40.0
79	Limestone, bluish to brown	0.2
78	Shale, buff	3.0
77	Limestone, dark gray, fine-grained, has unusually well preserved pectinoids, *Eumorphotis* cf. *iwanowi*	0.8
76	Shale, gray, silty	48.0
75	Limestone, dark gray, platy, fossils tend to weather in relief, poorly shown	4.0±
74	Shale, partially covered	33.0
73	Limestone, gray, fine-grained, poorly shown	0.5±
72	Shale, buff, with two or three thin blocky, fine-grained brown sandstones	38.5
71	Shale, with several thin gray limestones, fine-grained sandstones, poorly shown	75.0
70	Limestone, gray, fine-grained, brittle	2.5
69	Sandstone, brown, blocky, fine-grained, thin-bedded	14.0
68	Shale, buff, poorly exposed, contains some fine-grained thin-bedded blocky sandstone	41.0
67	Limestone, gray, weathers rough	2.5
66	Shale, covered	8.0
65	Sandstone, buff to brown, calcareous, fine-grained, contains *Lingula*, some shale partings	16.5
64	Shale, buff, silty	18.5
63	Limestone, brown, thin-bedded, alternating with buff siltstones, contains *Myalina*, shaly above	40.0
62	Shale and siltstone, contains a few poorly exposed brown thin limestones with *Lingula*	37.5
61	Limestone, brown, interbedded with buff siltstone	12.0
60	Shale, buff, silty, and buff siltstone	3.5
59	Limestone, bluish to brown, with shale partings	2.0
58	Shale, buff, silty, thin-bedded limestone in middle	13.0
57	Limestone, bluish, thin-bedded, fine-grained	1.0
56	Shale, buff, silty	1.5
55	Limestone, gray, fine-grained, bluish here fresh	2.5
54	Shale, buff, silty, one or two beds of thin brown limestone	11.0
53	Limestone, brown, interbedded with buff siltstone	3.0
52	Shale, silty, with three or four beds of thin brown limestone	14.5
51	Limestone, dark brown, bluish here fresh	1.5
50	Shale, buff, silty	8.0
49	Limestone, bluish, weathers brown	1.5
48	Shale, buff, silty	11.0
47	Limestone and interbedded siltstone	3.0
46	Shale, buff, silty	12.5
45	Limestone, even-bedded, fine-grained	1.5
44	Shale, buff, silty	16.0
43	Limestone, blue, in two thin beds, separated by buff calcareous siltstone	2.0
42	Siltstone, buff, shaly	19.0
41	Limestone, gray to brownish	0.8
40	Shale, olive to buff	17.0

Bed No.		Feet
39	Limestone, bluish-gray, fine-grained, even-bedded	3.5
38	Shale, buff, silty	39.0
37	Limestone, brownish-gray, blocky	1.5
36	Shale, buff, silty	13.0
35	Limestone, bluish-gray, fine-grained, blocky	1.5
34	Shale, buff, silty	15.0
33	Limestone, bluish, blocky, fine-grained, even-bedded, divided into three layers with shale and siltstone partings	3.0
32	Shale, buff, with 2-inch limestone above middle	10.5
31	Limestone, bluish, fine-grained	1.0
30	Siltstone, olive to buff, banded, platy	8.0
29	Limestone, brown	0.2
28	Shale, buff, silty	24.2
27	Limestone, bluish, weathers rounded, buff, thin-bedded	4.0
26	Shale, buff, silty, contains a few beds of hard platy siltstone	32.0
25	Limestone, bluish-gray, brittle	2.0
24	Shale, buff, silty, mainly covered	35.0
23	Limestone, blue, fine-grained, alternating with buff calcareous siltstone	4.0
22	Shale, buff, silty, contains some beds of hard calcareous buff siltstone	19.5
21	Limestone, bluish, fine-grained, below, grades upward into buff calcareous siltstone	2.5
20	Shale, buff, silty, platy	37.5
19	Limestone, bluish, fine-grained, brittle, blocky	1.5
18	Siltstone, olive to dark buff, calcareous	15.0
17	Limestone, in thin 1-inch bands, separated by buff siltstone	1.5
16	Shale, buff, silty	22.5
15	Limestone, buff, silty below, grading up into gray, fine limestone above	2.0
14	Shale, buff, silty, not well exposed	75.0
13	Limestone, blue, fine-grained	1.0±
12	Shale, buff, silty, poorly exposed	17.5
11	Limestone, gray to brown, fine-grained, thin-bedded, even	3.0
10	Shale, buff, silty, with four or five 3-inch beds of silty buff limestone in lower third of member	19.0
9	Limestone, brown, interlayered with buff calcareous siltstone, thin-bedded, contains some *Lingula*	4.0
8	Shale, buff, silty	13.5
7	Limestone, brown, thin-bedded, bluish where fresh	1.5
6	Shale, buff, silty, poorly exposed	9.8
5	Limestone, blue, fine-grained, thin-bedded, weathers light brown	3.5
4	Shale, buff, calcareous, platy, with a 3-foot buff limestone near base	6.0
3	Limestone, bluish-gray, fine-grained, even-bedded, weathers brown, contains a few *Lingula* and pelecypods	5.0
2	Shale and siltstone, buff to olive, mainly covered	350.0
1	Shale, brown, mostly covered, some platy brown siltstone	100.0
	Total Woodside and Dinwoody	2443.5

Phosphoria formation

Rex chert, dark gray to black.

22.—Red Canyon, Wind River Mountains, Fremont County, Wyoming
T. 5N., R. 6W.

Bed No.		Feet
	Red Peak formation (section incomplete)	
74	Sandstone, brick-red, slabby, with numerous thin shale partings, makes a bench	40.0
73	Shale, dark brick-red, silty, contains a few platy siltstones	11.0

Bed No.		Feet
72	Sandstone, red, platy and shaly, fine-grained, light gray streak at base	6.0
71	Siltstone or fine sandstone, rippled, platy, dark brick-red, silty and argillaceous shales	35.0
70	Siltstone, gray, shaly	1.0
69	Shale, dark brick-red, silty, poorly exposed	95.0
68	Shale, gray, silty	1.0
67	Shale, dark brick-red, silty	8.0
66	Siltstone, gray and red, interbedded with silty shales, somewhat calcareous bed at top	12.0
65	Shale, dark brick-red, silty, contains a few gray shale bands	30.0
64	Shale, buff, silty, grayish toward top	5.0
63	Shale, buff, sandy, soft, contains some massive units	3.0
62	Shale, lower part brown, middle gray, upper brown, more massive, weathered toward top	35.0
61	Shale, grayish-brown, hard	5.0
60	Shale, sandy and silty, makes obscure ledge	5.0
59	Shale, bluish-gray	104.0
58	Siltstone, gray, limy, thin-bedded, weathering buff	0.5
57	Shale, like 55	6.0
56	Shale, silty, hard	0.9
55	Shale, gray, soft	3.0
54	Siltstone, limy, thin-bedded	0.4
53	Shale, gray	11.0
52	Shale, buff, silty, contains fine-grained limy beds 1 inch thick	10.0
		427.8
	Dinwoody formation	
51	Sandstone, grayish-white, soft, massive, weathers dark brown	12.0
50	Shale, gray, sandy	0.4
49	Limestone, gray, silty, thin-bedded	1.7
48	Shale, greenish-gray, silty	7.0
47	Covered	7.0
46	Siltstone, sandy, shaly, weathers dark brown, contains abundant small pelecypods	4.2
45	Shale, gray, sandy	2.5
44	Siltstone, light gray, weathers dark brown, contains some shale beds, some beds a coquina of small clams; *Claraia* and *Myalina* present	7.0
43	Siltstone, gray, massive, weathering to shaly structure	4.0
42	Shale, gray silty	0.6
41	Siltstone, gray, massive, hard	4.0
40	Shale, silty and limy, some units fairly massive, weathers to paper-thin beds	4.0
39	Shale, gray, limy, weathers to paper-thin beds	0.6
38	Limestone, gray, silty, fossiliferous, weathers to paper-thin beds	0.4
37	Shale, greenish-gray, thin-bedded	2.5
36	Sandstone, fine-grained, lower part calcareous, massive, fossiliferous	2.5
35	Shale, gray, silty	1.5
34	Sandstone, fine-grained, lower part calcareous, massive, fossiliferous	2.5
33	Covered	2.0
32	Sandstone, fine-grained, lower part calcareous, massive, fossiliferous	4.5
31	Covered	4.0
30	Sandstone, light buff, massive, weathers dark brown	4.5
29	Shale, buff, sandy	1.5
28	Siltstone, buff, sandy in parts, irregularly bedded, contains some shaly units	14.0
27	Sandstone, fine-grained, massive, contains small pelecypods	3.0
26	Shale, buff, silty, contains some massive units	4.5
25	Siltstone, massive, sandy, weathers dark brown, contains small pelecypods	3.0
24	Shale, gray, silty	0.9
23	Siltstone, buff, sandy, massive	2.0

Bed No.		Feet
22	Shale, buff, silty	0.6
21	Siltstone, buff, massive, thinner bedded and more sandy near top, contains small pelecypods	1.5
20	Shale, gray, weathers buff, silty, contains small pelecypods	7.5
19	Sandstone, buff, weathers dark brown	2.0
18	Covered	5.0
17	Limestone, same as 15	1.0
16	Shale, gray, silty	0.3
15	Limestone, gray, weathers dark brown, has calcite geodes	1.3
14	Shale, gray, silty, weathering dark brown, contains small pelecypods	9.0
13	Siltstone, calcareous, weathers dark brown, contains small pelecypods	1.5
12	Shale, gray, silty	1.5
11	Siltstone, gray, weathers dark brown, massive, contains small pelecypods	1.0
10	Shale, gray, silty toward top	0.9
9	Sandstone, grayish-brown, silty	0.2
8	Limestone, same as 6	3.0
7	Shale, gray, soft	1.3
6	Limestone, impure, weathers dark brown, contains *Lingula* and small pelecypods	3.0
5	Shale, gray, silty, contains a few siltstones	1.4
4	Siltstone, gray, massive, thin-bedded toward top, weathers dark brown	1.5
3	Shale and siltstone, weathers dark brown	1.0
2	Shale	3.0
1	Clay, gray, soft, partially indurated, contains iron stain spots	0.7
	Total Dinwoody	156.5

Phosphoria formation

23.—Devils Slide, North of Gardiner, Cinnabar Mountains, Montana

Bed No.		Feet
	Red Peak formation (section incomplete)	
33	Sandstone, brown, gray, hard, platy, fine-grained, probably no red, several thin beds of gypsum, some thin blue silty shales	45.0
32	Sandstone, gray to olive, platy, fine-grained, wavy, rippled, grades laterally into red	10.0
31	Shale, red, silty, hard	3.5
30	Limestone or dolomite, white and pink	0.5
29	Shale, transition variable maroon and red, platy, silty	1.0
28	Shale, red, platy, hard, silty	3.5
27	Shale, greenish gray, platy	0.3
26	Shale, red, hard, platy, silty	7.0
25	Shale, red, platy, hard, gypsum in two or three beds	2.0
24	Shale, red and gray, platy	1.0
23	Sandstone, red, hard, fine-grained, platy	7.0
22	Gypsum	0.2
21	Sandstone, red, hard, fine-grained platy	4.5
20	Sandstone, whitish to gray, fine-grained, platy	1.0
19	Sandstone, brick-red, fine-grained, platy, 0.3 foot platy gray shale at middle	7.5
18	Shale, maroon, calcareous, platy	1.0
		95.0
	Dinwoody formation	
17	Limestone, light gray, dense, laminated or varved, platy, contains a few tiny high-spired unornamented gastropods	1.0
16	Shale and siltstone, gray, platy, sandier and olive toward base	2.0
15	Sandstone, olive, fine-grained, with gypsum veins	5.0

Bed No.		Feet
14	Siltstone or silty shale, olive, probably some very thin gypsum beds	19.0
13	Gypsum, white, bedded	0.25
12	Shale, olive, silty, with thin gypsum beds at middle	5.0
11	Gypsum, white, bedded	0.3
10	Shale, olive, silty	3.75
9	Gypsum bed	0.25
8	Shale, olive, gypsiferous	4.0
7	Dolomite, light-brown, silty	0.5
6	Shale, olive, silty, less than 1 inch thin gypsum	14.0
5	Sandstone, buff to gray, fine-grained, hard	1.0
4	Gypsum, white	1.0 to 1.5
3	Sandstone, light-gray, platy, gypsiferous, with interbedded gypsum	3.5
2	Shale, olive to gray clay, platy, with carbonaceous "spots"	1.0
1	Siltstone, gray, platy, calcareous	0.5
	Total Dinwoody	62.05 to 62.55

Phosphoria formation

Limestone, dark gray to black, bituminous, oölitic, platy

REFERENCES CITED

Baker, A. A., and Williams, J. S. (1940) *Permian in parts of Rocky Mountains and Colorado Plateau region,* Am. Assoc. Petrol. Geol., Bull., vol. 4, p. 617-636.

Bittner, A. (1899) *Versteinerungen aus den Trias-Ablagerungen des Süd Ussuri Gebietes,* Mem. Com. Geol., St. Petersburg, vol. 7, p. 1-35.

Blackwelder, E. (1918) *New geological formations in western Wyoming,* Washington Acad. Sci., Jour., vol. 8, p. 417-426.

Boutwell, J. M. (1907) *Stratigraphy and structure of the Park City mining district, Utah,* Jour. Geol., vol. 15, p. 434-458.

——— (1912) *Geology and ore deposits of the Park City district, Utah,* U. S. Geol Survey, Prof. Paper 77, p. 1-231.

Branson, E. B., and Branson, C. C. (1941) *Geology of Wind River Mountains, Wyoming,* Am. Assoc. Petrol. Geol., Bull., vol. 25, p. 120-151.

Diener, Carl (1895) *Triadische Cephalopodenfaunen der Ostsibirischen Küstenprovinz,* Mem. Com. Geol., St. Petersburg, vol. 14, p. 1-59.

Hyatt, Alpheus, and Smith, J. P. (1905) *The Triassic cephalopod genera of America,* U. S. Geol. Survey, Prof. Paper 40, p. 1-394.

King, R. E. (1940) *The geology of the Glass Mountains, Texas,* Part II, Texas Univ., Bull. 3042, p. 1-245.

Kiparisova, L. (1938) *The Lower Triassic pelecypods of the Ussuriland,* Travaux de l'Institut géologique de l'Académie des sciences de l'U. R. S. S., tome 7, p. 197-311.

Longwell, C. R., and Dunbar, C. O. (1936) *Problems of Pennsylvanian-Permian boundary in southern Nevada,* Am. Assoc. Petrol. Geol., Bull., vol. 20, p. 1198-1207.

Love, John D. (1939) *Geology along the southern margin of the Absaroka Range, Wyoming,* Geol. Soc. Am., Spec. Papers, no. 20.

Mansfield, G. R. (1927) *Geography, geology and mineral resources of part of southeastern Idaho,* U. S. Geol. Survey, Prof. Paper 152, p. 1-453.

McKee, E. D. (1938) *The environment and history of the Toroweap and Kaibab formation of northern Arizona and southern Utah,* Carnegie Inst. Washington, Publ. 492, p. 1-268.

Miller, A. K., and Cline, L. M. (1934) *The cephalopods of the Phosphoria formation of northwestern United States,* Jour. Paleont., vol. 8, p. 281-302.

———, and Furnish, W. M. (1940) *Permian ammonoids of the Guadalupe Mountains region and adjacent areas,* Geol. Soc. Am., Spec. Papers, no. 26, p. 1-242.

Muller, S. W., and Ferguson, H. G. (1939) *Mesozoic stratigraphy of the Hawthorne Tonopah quadrangles, Nevada,* Geol. Soc. Am., Bull., vol. 50, p. 1573-1624.

Newell, N. D., and Kummel, B. (1941) *Permo-Triassic boundary in Idaho, Montana and Wyoming,* Am. Jour. Sci., vol. 239, p. 204-208.

Smith, J. P. (1932) *Lower Triassic ammonoids of North America,* U. S. Geol. Survey, Prof. Paper 167, p. 1-199.

Spath, L. F. (1930) *The Eo-Triassic invertebrate fauna of East Greenland,* Meddelelser om Grønland, vol. 83, p. 1-87.

——— (1935) *Additions to the Eo-Triassic invertebrate fauna of East Greenland,* Meddelelser om Grønland, vol. 98, no. 2, p. 1-115.

Stewart, W. S. (1939) *"Park City" beds on southwest flank of Uinta Mountains, Utah,* Am. Assoc. Petrol. Geol., Bull., vol. 23, p. 82-100.

Thomas, H. D. (1934) *Phosphoria and Dinwoody tongues in lower Chugwater of central and southeastern Wyoming,* Am. Assoc. Petrol. Geol., Bull., vol. 18, p. 1655-1697.

Wilson, C. W. (1934) *Section of Paleozoic and Mesozoic rocks at Cinnabar Mountain, Park County, Montana, and at Mount Everts, Yellowstone National Park, Wyoming,* Am. Assoc. Petrol. Geol., Bull., vol. 18, p. 368-379.

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Explanation of Plates

PLATE 2.—CHARACTERISTIC DINWOODY FOSSILS

Figures 1–4. *Lingula borealis* Bittner (p. 953)

(1) External view of *Lingula* valve, hypotype ×4; 21351 Geol. Mus., Univ. Wis., from bed 13, Dinwoody formation, Martin Creek, Wyoming, locality 13.

(2) Internal valve, hypotype ×4; 21353 Geol. Mus., Univ. Wis., Dinwoody formation, western Wyoming.

(3) Internal valve, hypotype ×4; 21350 Geol. Mus., Univ. Wis., from bed 14, Dinwoody formation, Bear Gulch, Hogback Mountains, Wyoming, locality 11.

(4) External view of *Lingula* valve, hypotype ×4; 21352 Geol. Mus., Univ. Wis., from *Lingula* zone, Dinwoody formation, western Wyoming.

Figures 5a–b. *Terebratula margaritowi* Bittner (p. 954)

Dorsal and ventral valves, respectively, of typical specimens, hypotype ×2; 21340 Geol. Mus., Univ. Wis., from bed 105, Dinwoody formation, Henry, Idaho, locality 21.

Figure 6. *Discophiceras subkyokticum* Spatch (p. 959)

This is the best specimen of several ammonoids found in the Dinwoody formation; hypotype ×2; 21340 Geol. Mus., Univ. Wis., from a bed 150 feet above the base of the Dinwoody formation at Melrose, Montana, locality 20.

Figure 7. *Metophiceras subdemissum* Spath (p. 960)

Poorly preserved specimen, showing, however, the essential characters of Spath's species; hypotype ×2; 21348 Geol. Mus., Univ. Wis., from bed 12, Dinwoody formation, Montpelier Canyon, Idaho, locality 19.

Figure 8. *Mentzelia* sp.? (p. 954)

Rubber cast of ventral valve, hypotype ×4; 21339 Geol. Mus., Univ. Wis., from bed 40, *Lingula* zone, Dinwoody formation, MacDougals Pass, Salt River Range, Wyoming, locality 15.

Figure 9. *Spiriferina mansfieldi* Girty (p. 953)

Rubber cast of an incomplete ventral valve; hypotype ×4; 21337 Geol. Mus., Univ. Wis., from bed 2, *Lingula* zone, Green River Lakes, Wyoming, locality 9.

Figures 10–11. *Eumorphotis multiformis* Bittner (p. 957)

(10) Well-preserved left valve; hypotype ×1; 21363 Geol. Mus., Univ. Wis., from bed 2, *Claraia* zone, Dinwoody formation, Melrose, Montana, locality 20.

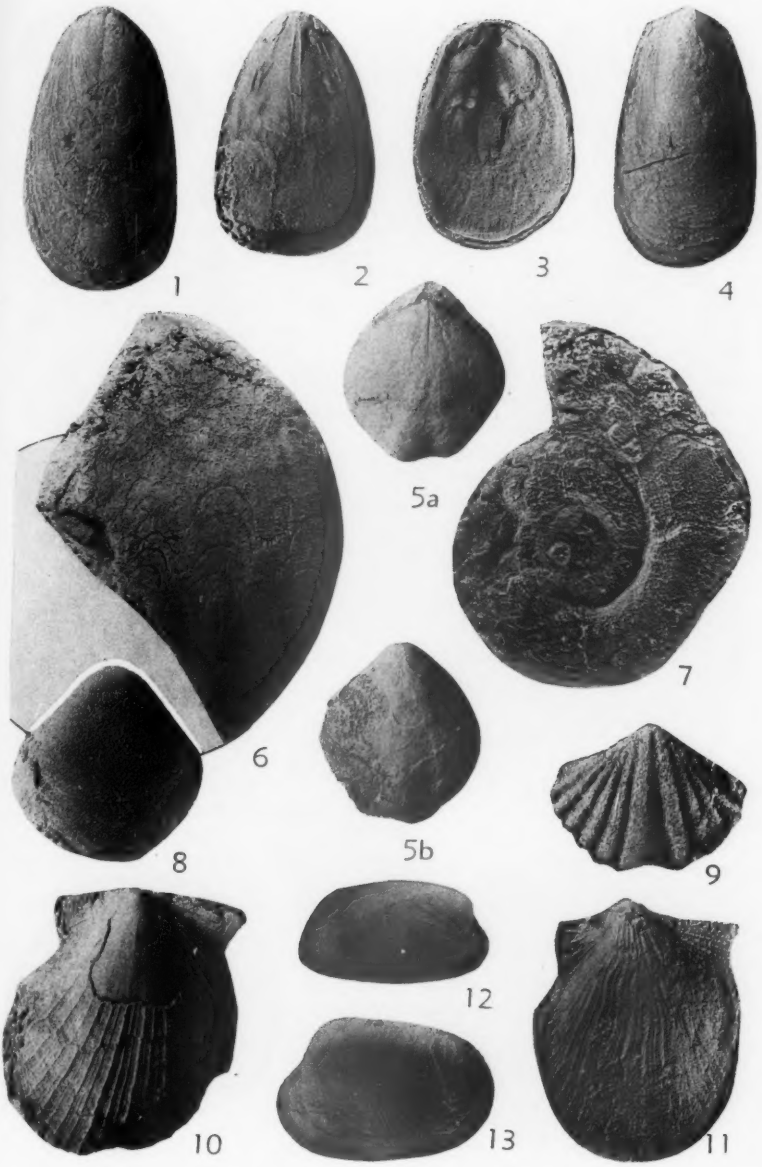
(11) Left valve, hypotype ×2; 21340 Geol. Mus., Univ. Wis., from bed 79, *Claraia* zone, Dinwoody formation, Montpelier Canyon, Idaho, locality 19.

Figure 12. *Pleurophorus* ? *bregeri* Girty (p. 957)

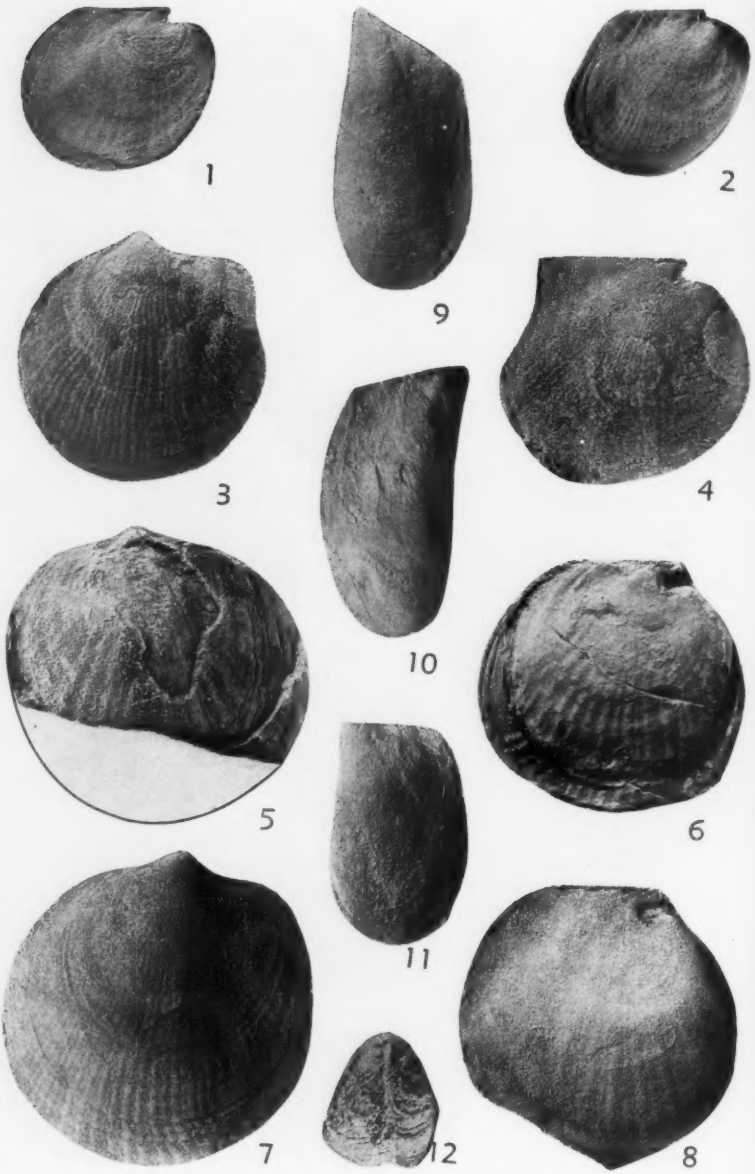
Right valve, hypotype ×2; 21362 Geol. Mus., Univ. Wis., from bed 28, *Claraia* zone, Dinwoody formation, Montpelier Canyon, Idaho, locality 19.

Figure 13. *Anodontophora fassaensis* (Wissmann) (p. 958)

Left valve, hypotype ×2; 21356 Geol. Mus., Univ. of Wis., from top limestone, *Claraia* zone, Dinwoody formation, South Fork Canyon, Wind River Mountains, Wyoming, N.W. ¼ sec. 17, T.1 S., R.2 W., locality 3.



CHARACTERISTIC DINWOODY FOSSILS



CHARACTERISTIC DINWOODY FOSSILS

PLATE 3.—CHARACTERISTIC DINWOODY FOSSILS

Figures 1–2. *Claraia clarai* var. *occidentalis* n. var. (p. 955)

(1) Right valve, holotype, ×1; 21343 Geol. Mus., Univ. Wis., from lower part of *Claraia* zone, Dinwoody formation, Meadow Creek Canyon, Wind River Mountains, Wyoming, sec. 2, T. 3 N., R. 5 W.

(2) Right valve slightly crushed on left side, paratype, ×1; 21342 Geol. Mus., Univ. Wis., from bed 18, *Claraia* zone, Dinwoody formation, Martin Creek, Wyoming Range, locality 13.

Figures 3–4. *Claraia mulleri* n. sp. (p. 956)

(3) Left valve, paratype ×1; 21358 Geol. Mus., Univ. Wis., from bed 7, *Claraia* zone, Dinwoody formation, Gros Ventre Canyon, Wyoming, locality 10.

(4) Right valve, holotype ×1; 21357 Geol. Mus., Univ. Wis., from same bed as specimen above.

Figures 5–8. *Claraia stachei* Bittner (p. 955)

(5) Incomplete specimen of a left valve, hypotype ×2; 20868 Geol. Mus., Univ. Wis., from a bed 10 feet above the base of the Dinwoody formation, *Claraia* zone, hidden anticline, 12 miles southeast of Lander, Wyoming.

(6) Poorly preserved right valve, hypotype ×2; 20867 Geol. Mus., Univ. Wis., from same locality as specimen above.

(7) Rubber cast of a rather large left valve, hypotype X1; 21345 Geol. Mus., Univ. Wis., from bed 38 *Claraia* zone, Dinwoody formation, Martin Creek, Wyoming Range, locality 13.

(8) Right valve, hypotype X1; 21344 Geol. Mus., Univ. Wis., *Claraia* zone, western Wyoming.

Figures 9–10. *Myalina putiatinensis* Kiparisova (p. 957)

(9) Well preserved left valve with a steep sloping hinge line, hypotype X1; 21359 Geol. Mus., Univ. Wis., from bed 2 *Lingula* zone, Dinwoody formation, Green River Lakes, Wyoming, locality 9.

(10) Right valve with a lower hinge angle, hypotype X1; 21360 Geol. Mus., Univ. Wis., from bed 2, *Lingula* zone, Dinwoody formation, Gros Ventre Canyon, Wyoming, locality 10.

Figure 11. *Myalina spathi* n.sp. (p. 956)

Left valve of holotype X1; 20870 Geol. Mus., Univ. Wis., this species has a form similar to one figured by Spath from the *Otoceras* zone of East Greenland; from bed 9, *Claraia* zone, Dinwoody formation, Melrose, Montana, locality 20.

Figure 12. *Bellerophon bittneri* n.sp. (p. 954)

Holotype X2; 21341 Geol. Mus., Univ. Wis., from a bed 10 feet above the base of the Dinwoody formation, *Claraia* zone, Red Canyon, Wind River Mountains, 12 miles southeast of Lander, Wyoming.

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